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Seafloor geomorphology and glacimarine sedimentation associated with fast-flowing ice sheet outlet glaciers in Disko Bay, West Greenland

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Abstract

Fast-flowing outlet glaciers currently drain the Greenland Ice Sheet (GIS), delivering ice, meltwater and debris to the fjords around Greenland. Although such glaciers strongly affect the ice sheet's mass balance, their glacimarine processes and associated products are still poorly understood. This study provides a detailed analysis of lithological and geophysical data from Disko Bay and the Vaigat Strait in central West Greenland. Disko Bay is strongly influenced by Jakobshavn Isbræ, Greenland's fastest-flowing glacier, which currently drains ~7% of the ice sheet. Streamlined glacial landforms record the former flow of an expanded Jakobshavn Isbræ and adjacent GIS outlets through Disko Bay

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and the Vaigat Strait towards the continental shelf. Thirteen vibrocores contain a complex set of lithofacies including diamict, stratified mud, interbedded mud and sand, and bioturbated mud deposited by (1) suspension settling from meltwater plumes and the water column, (2) sediment gravity flows, and (3) iceberg rafting and ploughing. The importance of meltwater-related processes to glacial marine sedimentation in West Greenland fjords and bays is emphasised by the abundance of mud preserved in the cores. Radiocarbon dates constrain the position of the ice margin during deglaciation, and suggest that Jakobshavn Isbræ had retreated into central Disko Bay before 10.6 cal ka BP and to beyond Isfjeldsbanen by 7.6–7.1 cal ka BP. Sediment accumulation rates were up to 1.7 cm a⁻¹ for ice-proximal glacial marine mud, and ~0.007–0.05 cm a⁻¹ for overlying distal sediments. In addition to elucidating the deglacial retreat history of Jakobshavn Isbræ, our findings show that the glacial marine sedimentary processes in West Greenland are similar to those in East Greenland, and that variability in such processes is more a function of time and glacier proximity than of geographic location and associated climatic regime.

Keywords: Glacial geomorphology, sedimentology, Holocene, Greenland, deglaciation, tidewater glaciers

1. Introduction

Tidewater glaciers terminate in the ocean at a grounded ice front (Meier & Post, 1987), represent an important link between terrestrial and marine environments, and are particularly susceptible to climate change. Along the coast of Greenland many fast-flowing outlet glaciers drain the interior of the Greenland Ice Sheet (GIS), terminating as tidewater margins in the surrounding fjords. The associated glacial landforms and glacial marine sediments are revealed as the glaciers retreat, and provide important archives for understanding the long-term glacial evolution of the ice sheet and its future role with respect to sea-level rise (cf. e.g. Alley et al., 2005; Bamber et al., 2007; Nick et al., 2009; Ó Cofaigh et al., 2013; Dowdeswell et al., 2014; Lane et al., 2014; Joughin et al., 2014;

12 Hogan et al., 2016; Sheldon et al., 2016). Jakobshavn Isbræ, in central West
 13 Greenland, is of particular interest in this context, as it is the fastest-flowing
 14 of these outlets, currently flowing at velocities $>17 \text{ km a}^{-1}$ and draining $\sim 7\%$
 15 of the GIS, and thus exerts a strong influence on the ice sheet’s mass balance
 16 (e.g. Bindshadler, 1984; Joughin et al., 2004; Rignot & Kanagaratnam, 2006;
 17 Joughin et al., 2014). Indeed, the increasing retreat speed and break-up of the
 18 glacier tongue has led to a rise in global sea level of almost 1 mm between
 19 2000 and 2011 (Howat et al., 2011; Joughin et al., 2014). Although a num-
 20 ber of investigations have focussed on the short-term dynamics of GIS outlet
 21 glaciers (e.g. Joughin et al., 2004; Moon & Joughin, 2008; Joughin et al., 2014),
 22 knowledge about their longer-term flow dynamics, their glacimarine processes,
 23 and the overall interaction of the glaciers with the marine environment since the
 24 Last Glacial Maximum (LGM) is only just emerging (e.g. Long & Roberts, 2003;
 25 Young et al., 2011a; Jennings et al., 2013; Ó Cofaigh et al., 2013; Dowdeswell
 26 et al., 2014; Hogan et al., 2016; Sheldon et al., 2016). This study uses sediment
 27 cores, multibeam bathymetry, sub-bottom profiler data, and radiocarbon dates
 28 from Disko Bay and the Vaigat Strait (Fig. 1) to (1) investigate the Holocene
 29 glacimarine sedimentary processes and products in Disko Bay and (2) to eluci-
 30 date the deglacial history of Jakobshavn Isbræ in order to see how this particular
 31 outlet responded to environmental changes since the LGM.

32 **2. Study area**

33 **2.1. Physiographic setting**

34 Disko Bay is a marine embayment in central West Greenland, which is separated
 35 from the Vaigat Strait, a relatively narrow deep water trough, by Disko Island
 36 (Fig. 1). Disko Bay is located between $\sim 68^{\circ}30' - 69^{\circ}40' \text{N}$ and $50^{\circ}50' - 55^{\circ}00' \text{W}$,
 37 and is roughly 100 km wide and between 50 and 500 m deep. It covers an area
 38 of $\sim 18000 \text{ km}^2$ and is bounded by Isfjorden and the Greenland mainland to the
 39 east, and Baffin Bay to the west (Fig. 1). A large, relatively shallow ridge,

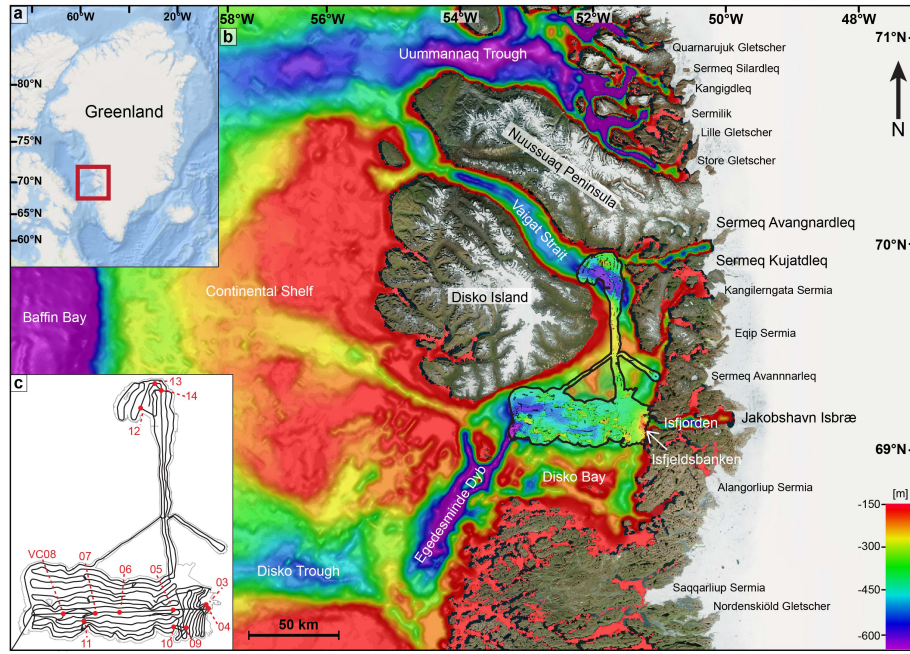


Figure 1: a) Overview of Greenland with red rectangle indicating the extent of b). b) Study area and local bathymetry (IBCAO). The black outline shows the extent of the bathymetric data available for this study. Purple areas indicate bathymetric troughs. c) Distribution of TOPAS lines and location of the vibrocores.

Isfjeldsbanken, is located at the entrance of the smaller Isfjorden and serves as a sill between the latter and Disko Bay. The Vaigat Strait is situated between $\sim 69^{\circ}40' - 70^{\circ}50'N$ and $50^{\circ}50' - 55^{\circ}00'W$ (Fig. 1), is 10–30 km wide, and 200–650 m deep. It is bounded by the Nuussuaq Peninsula to the north and east and Disko Island to the south and west (Fig. 1). Three larger basins are present in the study area, one in the Vaigat Strait (up to 650 m deep) and two in western Disko Bay. The latter are ~ 800 m deep and part of Egedesminde Dyb, a large trough, which is orientated northeast-southwest linking Disko Bay with the continental shelf (Fig. 1b). The local geology is dominated by Precambrian basement, including crystalline rocks such as granites and orthogneisses along the western shore of the Greenland mainland, Palaeogene basalts on Disko Island and western Nuussuaq, and Palaeogene and Upper Cretaceous sediments

52 exposed at the seafloor and on parts of Disko Island and the Nuussuaq Peninsula
53 (Chalmers et al., 1999; Larsen & Pulvertaft, 2000; Weidick & Bennike, 2007).

54 **2.2. Glacial background**

55 Although there are still gaps in our understanding of the long-term evolution
56 of the GIS and its outlet glaciers (cf. Funder et al., 2011), recent studies have
57 outlined the Pliocene-Pleistocene glacial development of the Disko Bay margin
58 (Hofmann et al., 2016), and established that during the LGM an extended
59 Jakobshavn Isbræ and several other glaciers in the area drained the GIS via
60 Disko Bay and the Vaigat Strait, and extended to the outer shelf edge (Ó Cofaigh
61 et al., 2013; Jennings et al., 2013; Hogan et al., 2016). Radiocarbon dates from
62 reworked shells from the Disko trough-mouth fan and tills on the adjoining shelf
63 suggest that retreat of Jakobshavn Isbræ was underway by at least 13.8 ka BP
64 and was briefly interrupted around 12.3–12 ka BP when the ice sheet underwent
65 a re-advance in Disko Trough during the Younger Dryas (Ó Cofaigh et al., 2013).
66 Two modes of ice retreat have been suggested, (1) fast and relatively continuous
67 retreat from the continental shelf and through Disko Bay (e.g. Long & Roberts,
68 2003; Lloyd et al., 2005; Hogan et al., 2012; Kelley et al., 2013), and (2) step-
69 wise retreat, where Jakobshavn Isbræ experienced short periods of still-stand
70 at bedrock highs (e.g. Weidick, 1996; Rasch, 2000; Hogan et al., 2016). This led
71 to the general consensus that retreat across the continental shelf and through
72 Disko Bay was relatively fast, but slowed once the ice stream entered Isfjorden
73 in the east (cf. Funder & Hansen, 1996; Lloyd et al., 2005; Hogan et al., 2012;
74 Kelley et al., 2013; Ó Cofaigh et al., 2013).

75 Deglaciation of western Disko Bay commenced around 10.8 ka BP, and the
76 bay’s eastern part was ice-free by 10.2 ka BP (Lloyd et al., 2005; Kelley et al.,
77 2013). The grounded margin of Jakobshavn Isbræ most likely reached Isfjelds-
78 banken in eastern Disko Bay around 10.1 ka BP, pausing there until ~7.9 ka
79 BP, when it retreated into Isfjorden (see Fig. 1b; Lloyd et al., 2005; Weidick
80 & Bennike, 2007; Kelley et al., 2013). At present the Jakobshavn Isbræ margin

81 is located approximately 50 km east of Isfjeldsbanken and discharges 90–100
82 $\text{km}^3 \text{ a}^{-1}$ of ice into Isfjorden (Joughin et al., 2014). Due to a shorter calving
83 line, the calving flux from Jakobshavn Isbræ was suggested to be significantly
84 reduced around 9–10 ka BP, when the glacier margin was at Isfjeldsbanken, and
85 the glaciers in northeast Disko Bay were inferred to be the dominant source of
86 ice-rafted debris (IRD) during this time (Weidick, 1994; Long & Roberts, 2003;
87 McCarthy, 2011). Retreat of the outer parts of the GIS outlets was asynchronous
88 along Greenland’s western coast (Ó Cofaigh et al., 2013; Sheldon et al., 2016),
89 and deglaciation in the Vaigat was underway by 12.4 ka BP, with its western
90 part ice-free before 11.8 ka BP and its eastern part deglaciated before 10.0 ka
91 BP (Weidick, 1968; Bennike, 2000).

92 **2.3. Oceanography**

93 During deglaciation and the early Holocene, ocean waters in Disko Bay and the
94 Vaigat Strait were mainly dominated by cold and fresh meltwater from the GIS
95 (e.g. Lloyd et al., 2005; Jennings et al., 2013). By approximately 10 ka BP,
96 the West Greenland Current (WGC) started to bring warmer and more saline
97 waters into the bay, influencing the coastal areas around 7.8 ka BP, when ice had
98 retreated into Isfjorden and the meltwater flux into Disko Bay had decreased
99 (Lloyd et al., 2005; Lloyd, 2006). After c. 6 ka BP the regional circulation
100 pattern started to resemble modern conditions (Perner et al., 2013), and today
101 the modern tidewater glaciers still influence the surface waters in Disko Bay
102 and the Vaigat Strait, which are cold and fresh (Andersen, 1981; Ribergaard
103 & Buch, 2008). The bottom waters, however, contain warmer and more saline
104 waters from the WGC (Lloyd et al., 2005; Perner et al., 2013). These waters
105 are advected through Disko Bay from west to east and flow northwards around
106 Disko Island and through the Vaigat Strait (e.g. Andresen et al., 2010). They
107 not only influence iceberg calving rates, but have also been linked to increased
108 thinning and melting of GIS outlet glaciers (Holland et al., 2008; Rignot et al.,
109 2010; Kelley et al., 2013).

110 2.4. Acoustic stratigraphy of marine sediments

111 The sub-bottom profiler data available for this study were previously described
112 and interpreted by Hogan et al. (2011, 2012), who identified four acoustic facies
113 in Disko Bay, AD1–AD4, culminating in a total maximum thickness of up to
114 258 ± 8 m (calculated using a p-wave velocity of 1610 m s^{-1} ; Fig. 2a, b).
115 Facies AD1, with a stratified acoustic signature and a strong upper reflection
116 (Fig. 2a), is 16–64 m thick, has onlap-fill geometry, and forms wedges in places
117 (Hogan et al., 2012). Facies AD2 generally overlies and locally cuts into AD1,
118 is composed of acoustically transparent sub-units, and shows tapered or wedge-
119 shaped geometry. It is 4–32 m thick and its upper boundary generally occurs as
120 a continuous reflection of high amplitude (Fig. 2a; Hogan et al., 2012). Facies
121 AD3, like AD1, is acoustically stratified with internal reflections of medium
122 strength (Fig. 2a). AD3 conformably overlies AD2, drapes some of the bedrock
123 highs in the area and is up to 13 m thick. Facies AD4 only occurs in parts
124 of Disko Bay, where it appears acoustically transparent with weak and chaotic
125 internal reflections protruding into AD1 and AD2, and a strong, hummocky and
126 chaotic upper boundary (Fig. 2a; Hogan et al., 2012).

127 In southern Vaigat, Hogan et al. (2012) distinguished a total of five acoustic
128 facies, AV1–AV5, with a cumulative thickness of up to 109 ± 3 m (Fig. 2c, d).
129 Facies AV1, AV2, AV3, and AV4 are acoustically homogeneous with generally
130 weak, discontinuous to chaotic internal reflections and are bounded by medium-
131 strong, mostly continuous upper, in places hummocky, reflections. A distinction
132 into four acoustic facies was mostly based on different morphologies; while AV1
133 represents the deepest basin-infill strata in the Vaigat, AV2 has a distinct wedge-
134 shape, AV3 occurs as lenticular bodies, and AV4 infills surface depressions of
135 AV2 and AV3 (Fig. 2c, d).

136 Facies AD1 was inferred to contain sediment deposited from turbid meltwa-
137 ter plumes, from the water column, icebergs, and sediment gravity flows in an
138 ice-proximal environment in the eastern bay and in an ice-distal environment
139 in the western bay (Hogan et al., 2011, 2012). From the tapered/wedge-shaped

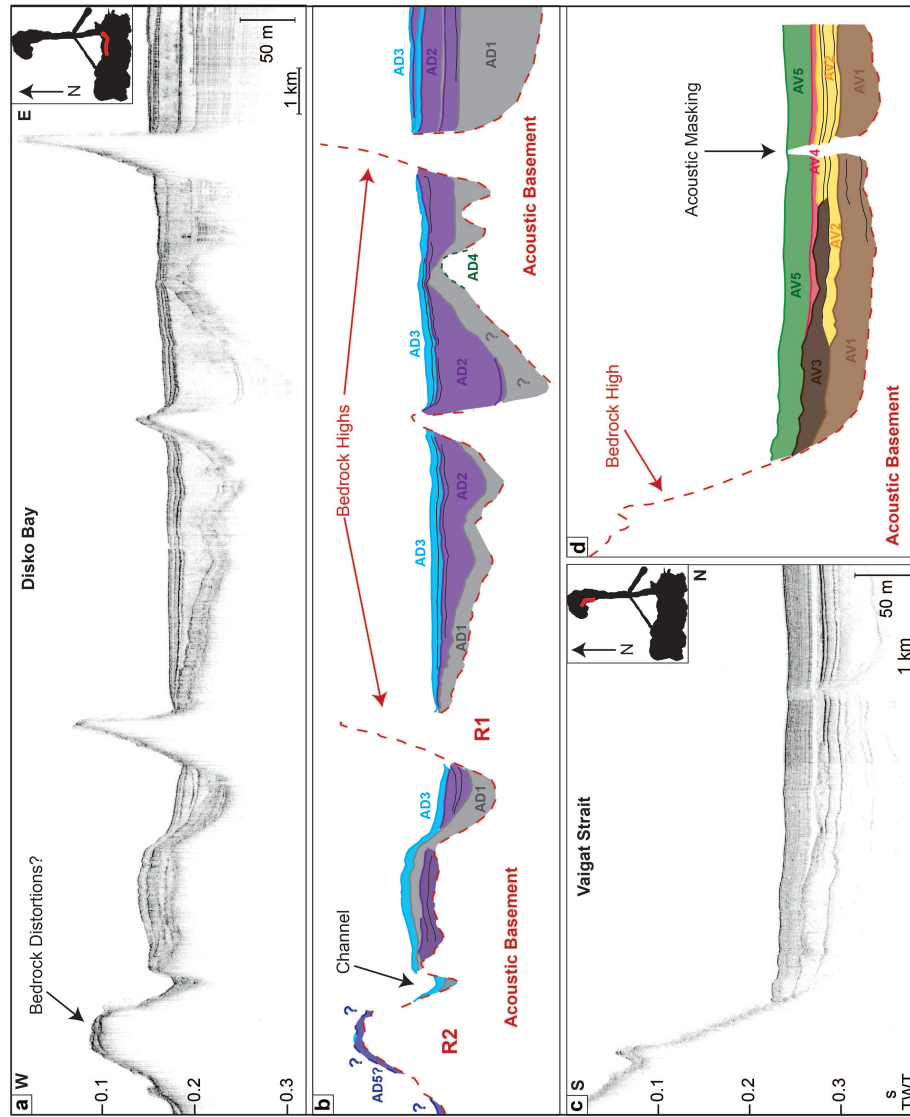


Figure 2: a) TOPAS profile showing an example of the acoustic facies in Disko Bay. b) Interpretation of acoustic facies in Disko Bay, after Hogan et al. (2012). c) TOPAS profile with examples of acoustic facies occurring in the Vaigat Strait. d) Acoustic facies interpretation in the Vaigat Strait based on Hogan et al. (2012). The red lines on the black polygons indicate the respective location of the profiles.

140 geometry and the acoustic transparency, the sub-units of AD2 were interpreted

141 to also reflect gravity-flow deposits. These are occasionally interbedded with
 142 thin sediment strata derived from hemipelagic sedimentation, bottom currents,
 143 and smaller-scale or more dilute gravity flows (Hogan et al., 2012). Hogan
 144 et al. (2011, 2012) interpreted Facies AD3 as ice-distal sediments settling from
 145 hemipelagic sedimentation and icebergs and/or sea ice. The internal reflections
 146 were associated with variations in input of IRD and/or bottom current activity.
 147 Facies AD4 was interpreted as a facies representing the upward migration of
 148 fluids through the sediment column (Hogan et al., 2012).

149 Facies AV1–AV4 in the Vaigat were interpreted as partly erosive gravity-
 150 flow deposits derived from: (i) the deposition and remobilisation of glacimarine
 151 sediment settling from turbid meltwater plumes in the case of AV1 and AV4;
 152 (ii) an interplay of suspension settling and bottom currents in the case of AV2;
 153 and/or (iii) slumps down bedrock slopes in the case of AV3 (Hogan et al., 2012).
 154 Facies AV5 forms a conformable drape over the existing topography and was
 155 inferred to be deposited by post-glacial hemipelagic sedimentation with variable
 156 input of IRD by icebergs and sea ice (Hogan et al., 2012).

157 3. Materials and methods

158 Nine vibrocores (VC03–VC11) from Disko Bay and three from the Vaigat (VC12–
 159 VC14; Fig. 1c) were collected in August 2009 during cruise JR175 of the *RRS*
 160 *James Clark Ross* to the West Greenland continental margin. Together with
 161 swath-bathymetric data (Fig. 1b), these sediment cores provide the basis for
 162 this study. The cores were acquired using the British Geological Survey vi-
 163 brocorer with a 6 m-long barrel and an inner diameter of approximately 9 cm.
 164 Core recovery was excellent in soft sediments to moderate in diamicts. Upon
 165 retrieval, all sediment cores were divided into ~1 m long sections, split into
 166 working and archive halves, and stored at +4°C. Core locations and lengths are
 167 summarised in Table 1. In order to identify the lithofacies, core logs of all work-
 168 ing halves were generated from the core sections, and x-radiographs were used to
 169 provide supplementary information on sub-surface sedimentary structures and

quantification of clasts larger than 2 mm, classified as IRD (sensu Grobe, 1987). A GEOTEK multi-sensor core logger (MSCL) was used to measure physical properties such as wet-bulk density, p-wave velocity (only VC05, VC07, VC09) and magnetic susceptibility (MS), which was acquired with a Bartington point-sensor mounted on the GEOTEK system. Shear strength measurements were undertaken with a Durham Geo Slope Indicator torvane and, for most cores, were carried out directly after splitting in 2009. For VC03 and VC04, however, the shear strength was only determined in 2016; hence values for these cores should be treated as estimates. Grain-size distribution and water content were measured by sampling approximately 1 cm-thick sediment slices in 8 cm-intervals, which were weighed, dried at 60°C and subsequently weighed and sieved through mesh sizes of 63, 125, 250, and 500 μm .

Samples for radiocarbon dating were collected from as close to distinct lithological boundaries in the cores as possible. Accelerator Mass Spectrometry (AMS) radiocarbon dates were measured at Beta Analytic on ~ 6 mg of mixed species benthic foraminifera, and additional radiocarbon dates were obtained from molluscs and seaweed at the INSTAAR NSRL laboratory. The conventional radiocarbon ages were calibrated into cal a BP using Calib 7.1 with the MARINE13 curve and a reservoir correction of $\Delta R = 140 \pm 25$ (Stuiver & Reimer, 1993; Lloyd et al., 2011; Reimer et al., 2013). The same calibration was applied to already published ^{14}C dates from marine shells in Disko Bay (Lloyd et al., 2005; McCarthy, 2011; Ó Cofaigh et al., 2013), in order to make dates directly comparable.

Swath-bathymetric data were acquired during the same cruise, using a hull-mounted Kongsberg Maritime Simrad EM120 multibeam echo sounder. The system operates at a frequency of 12 kHz and was calibrated using sound velocity profiles for the water column obtained from XBTs. The data were processed using MB-System Software and QPS Fledermaus, gridded to a cell size of 30x30 m in QPS DMagic, and visualised and interpreted in Fledermaus. The data were supplemented with swath-bathymetric data collected during two additional cruises to Disko Bay, one on *RV Maria S. Merian* in June 2007, and

Table 1: Core locations and recovery.

Core ID	Latitude	Longitude	Area	Depth [m]	Length [m]
VC03	69°10.81' N	51°11.61' W	Disko	545	1.57
VC04	69°09.97' N	51°10.15' W	Disko	263	1.10
VC05	69°09.60' N	51°31.63' W	Disko	389	5.87
VC06	69°08.94' N	52°04.14' W	Disko	439	4.94
VC07	69°08.62' N	52°18.88' W	Disko	439	5.46
VC08	69°08.35' N	52°38.24' W	Disko	429	3.91
VC09	69°05.79' N	51°23.65' W	Disko	294	5.98
VC10	69°05.95' N	51°31.22' W	Disko	351	4.86
VC11	69°06.90' N	52°25.60' W	Disko	410	3.25
VC12	69°53.12' N	51°53.15' W	Vaigat	616	3.66
VC13	69°58.46' N	51°44.47' W	Vaigat	341	3.40
VC14	69°56.97' N	51°40.35' W	Vaigat	386	4.66

201 the other on the private fishing vessel *MV Smilla* in August 2008. The *Merian*
 202 data were acquired with a hull-mounted Kongsberg Maritime EM120 multibeam
 203 echo sounder in deep water and a Kongsberg Maritime EM1002 in shallow wa-
 204 ter, with the former operating at a frequency of 12 kHz and the latter at 95 kHz.
 205 The data were processed in the MB-System software (sensu Caress & Chayes,
 206 1996) and gridded to a cell size of 24x24 m. The *MV Smilla* data were collected
 207 using a temporarily installed Sea Beam 1180 shallow water swath echo sounder
 208 system at a nominal frequency of 180 kHz, and gridded to a cell size of 15x15
 209 m. Sub-bottom profiler data are also available for this study (Fig. 1c) and
 210 were gathered simultaneously with the swath bathymetry from the 2009 *James*
 211 *Clark Ross* cruise, using a Kongsberg Maritime TOPAS PS18 sub-bottom pro-
 212 filer, which operated at a frequency of 3.5 kHz. These data were played out in
 213 near-real time with an EPC chart recorder installed on board the vessel, provid-
 214 ing high-resolution (30–40 cm) acoustic profiles, post-processed in the TOPAS
 215 Software, and subsequently loaded into IHS The Kingdom Software 2015. Con-
 216 version between milliseconds and metres was done using a p-wave velocity of
 217 1610 m s⁻¹, the combined average velocity measured in unconsolidated sedi-
 218 ments from the three cores from Disko Bay. The TOPAS data and parts of the
 219 swath-bathymetric data were already analysed and interpreted by Hogan et al.
 220 (2012) and TOPAS profiles are thus only used for correlation purposes in this
 221 study. Bathymetric data from the *Merian* and *Smilla* vessels were previously

222 interpreted by Schumann et al. (2012).

223 4. Results

224 4.1. Swath bathymetry

225 A geomorphological map of the landforms in Disko Bay and the Vaigat Strait
226 is shown in Figure 3. Earlier mapping from the easternmost part of Disko Bay
227 (Hogan et al., 2012; Schumann et al., 2012) is incorporated into this map.

228 **Large transverse ridges** The most prominent characteristic of the seafloor
229 is its rugged, irregular topography, imparted by a number of transverse ridges,
230 which are generally orientated in a north-south direction (Fig. 3). Most of these
231 ridges are relatively discontinuous and between 1 and 2 km long. They have
232 sharp crests imparted by steep eastern, and more gradual western flanks, the
233 majority of which are intensely streamlined in the direction of ice flow (generally
234 east-west). Three ridges, R1–R3, stand out morphologically (Fig. 3). R1 is the
235 most proximal ridge, concave in planform with respect to the ice margin, and
236 located approximately 20 km west of Isfjeldsbanken. It is ~ 4.5 km long, 40 m
237 high, and up to 500 m wide. R2, at 26 km from Isfjeldsbanken, is 20 km long,
238 200–1000 m wide, and 10–120 m high, with a generally convex crest forming
239 a slight zig-zag pattern (Fig. 4). The distal flanks of R1 and parts of R2 are
240 intensely streamlined (Fig. 4). R3 is curvilinear in plan view, 20 km long, up
241 to 4 km wide and 20–120 m high.

242 The large dimensions and the rugged appearance of R1–R3 indicate that
243 a purely glacial origin is unlikely (cf. Ottesen & Dowdeswell, 2006; Ottesen
244 et al., 2008; Hogan et al., 2011; Flink et al., 2015; Streuff et al., 2015), and the
245 sub-bottom profiler data show that the majority of the topographically distinct
246 highs are formed in bedrock (e.g. Fig. 4d). We therefore interpret these ridges
247 as bedrock highs that were overridden and streamlined by glacial ice.

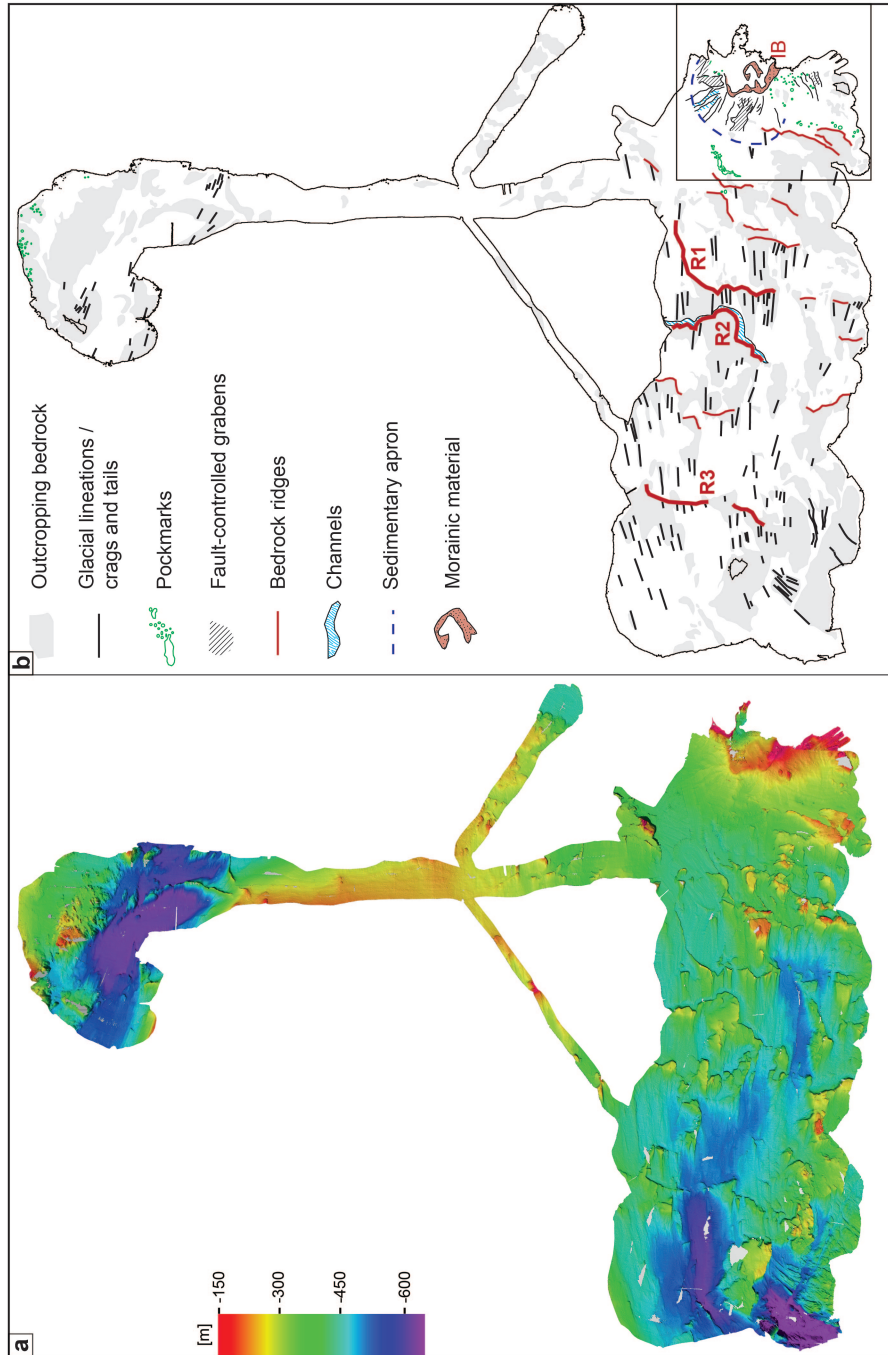


Figure 3: a) Bathymetry in Disko Bay and the Vaigat Strait. b) Geomorphological map of all the landforms in Disko Bay. Landforms in the black rectangle indicate those already mapped by Hogan et al. (2012) and Schumann et al. (2012). IB = Isfjeldsbanken. Detailed examples are shown in Fig. 4.

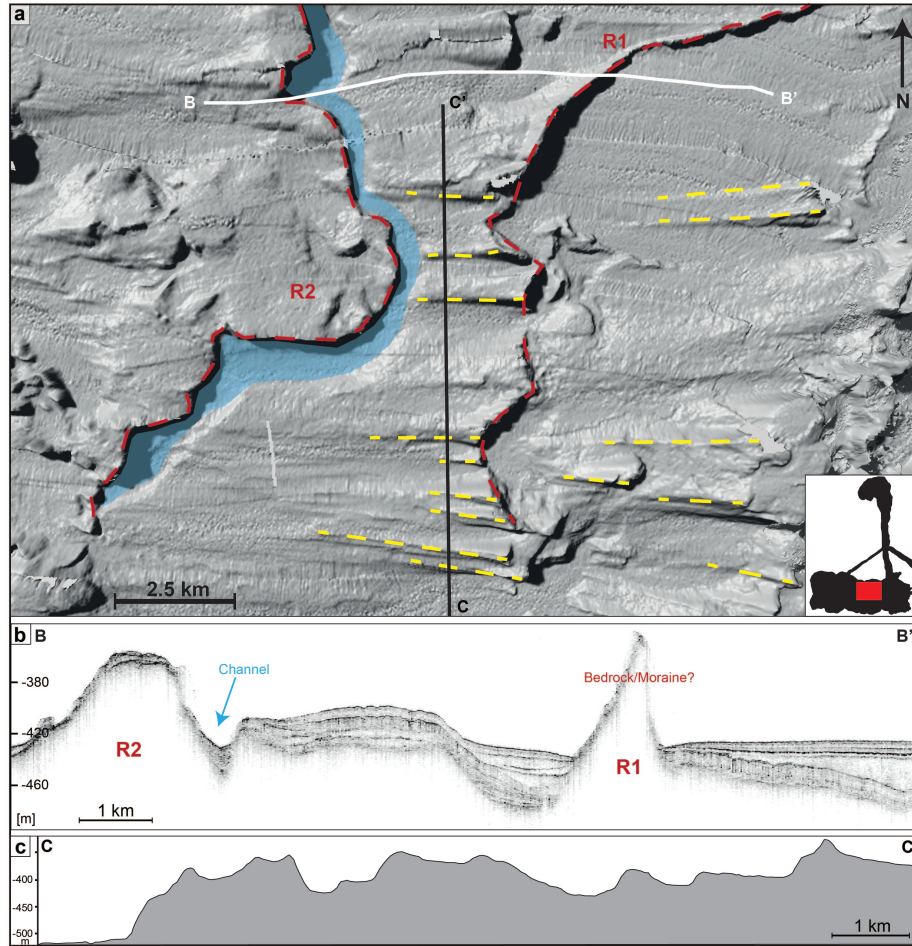


Figure 4: a) Shaded-relief image of the bathymetry in Disko Bay. Red lines show the location of bedrock ridges, while yellow stippled lines follow the long-axes of the crags and tails. The blue polygon shows the location and extent of C1, and the red rectangle on top of the black polygon in the bottom-right hand corner shows the extent and location of a). b) TOPAS profile BT-BT' across across a submarine channel and a ridge. c) Bathymetric profile C-C' across crag-and-tails.

248 **Elongate hills** The north-south orientated bedrock ridges in Disko Bay are
 249 closely associated with east-west orientated elongate hills (Figs. 3, 4). The lat-
 250 ter are 1.5–7 km long, 100–1000 m wide, and around 10 m high with typically
 251 broader, steeper stoss sides and gently tapering lee ends (Fig. 4). The char-

acteristics of these landforms are consistent with formation as crag-and-tails, the presence of which in Disko Bay was also documented by Ó Cofaigh et al. (2013). Crag-and-tails form subglacially and in association with bedrock highs, where the crag consists of bedrock with a lee-side tail forming from deposition of unconsolidated subglacial sediment (Dionne, 1987; Stokes et al., 2011).

Submarine channels A large channel, C1, occurs about 25 km west of Isfjeldsbanken and is ~16 km long, around 800 m wide and up to 40 m deep. It follows the eastern edge of R2 and is sinuous in planform (Fig. 4). Several similar, generally smaller depressions have also been observed along the western flank of Isfjeldsbanken (Fig. 3; Hogan et al., 2012). The large channel C1 is interpreted to be a subglacial channel eroded by meltwater flowing beneath an extended Jakobshavn Isbræ (cf. e.g. Walder & Hallet, 1979). The depth and shape of the channel imply that its formation took some time, during which meltwater erosion must have been focussed along R2. Assuming that meltwater disperses with increasing distance from the ice margin, concentrated meltwater routing implies that the ice margin was relatively close. As the channel is located on the proximal side and follows the line of R2, it seems plausible that the glacier front grounded on the bedrock high for an extended period of time and subglacial meltwater was routed around the bedrock obstacle eroding the channel. The smaller channels on the western flank of Isfjeldsbanken formed in the sediment pile and are interpreted as submarine channels eroded from downslope sediment-gravity flows, occasionally promoted by the presence of faults (Hogan et al., 2012).

Sediment-gravity flows A large sedimentary apron on the western flank of Isfjeldsbanken was described by Schumann et al. (2012) and several smaller incisions along the same flank were interpreted as sediment slumps from downslope-gravity flows (Hogan et al., 2012). Although such landforms do not always appear clearly on our bathymetric data, the sub-bottom profiler data indicate the abundance of such deposits in Disko Bay and the Vaigat Strait (see AD2 and

281 AV1–AV5 in Fig. 2). Common triggers of gravity-flows are, for example, contin-
282 uously high sediment accumulation and regional seismicity, the latter possibly
283 related to isostatic rebound (e.g. Hunt & Malin, 1998; Forwick & Vorren, 2012).

284 **Pockmarks** Several circular depressions occur in Disko Bay, and are especially
285 common in the eastern part of the bay and on the distal flank of Isfjeldsbanken
286 (Fig. 3; see also Figs. 6 and 8 in Hogan et al., 2012). The depressions often
287 occur in clusters, are between 5 and 300 m in diameter, and 7–30 m deep. On
288 the sub-bottom profiler data, the depressions are associated with a drawdown of
289 the overlying reflections and occasional acoustic masking (Hogan et al., 2012).
290 These depressions are interpreted as pockmarks (Hogan et al., 2012), which are
291 formed as a result of gas or pore fluid seepage (e.g. Harrington, 1985; Hovland &
292 Judd, 1988; Forwick et al., 2009; Nielsen et al., 2014; Dowdeswell et al., 2016).
293 Acoustic masking on the sub-bottom profiler data supports this interpretation.

294 4.2. Sub-bottom profiler data

295 **Description** Our seismostratigraphic findings support previous work from
296 Hogan et al. (2012), who identified four acoustic facies in Disko Bay, AD1–
297 AD4. Although difficult to discern, we identify one additional acoustic facies,
298 AD5, which conformably overlies and occasionally onlaps the acoustic basement
299 in localised areas of Disko Bay (see Figs. 2a, b, 5c, f). AD5 is characterised by
300 chaotic, semi-transparent internal reflections of variable strength and is 11 ms
301 (~ 9 m) at its thickest. It can be bounded by a strong upper reflector and can
302 appear slightly distorted by bedrock echos (Fig. 2a). AD5 differs from AD2 by a
303 slightly more opaque acoustic character with a larger number of internal reflec-
304 tions. Furthermore, unlike in AD2, the TOPAS signal weakens with increasing
305 depth and quickly disappears beneath the upper boundary of AD5.

306 **Interpretation** The semi-transparent and internally massive acoustic appear-
307 ance of Facies AD5 as well as a decreasing signal strength with depth have
308 sometimes been attributed to uniformly mixed sediments of possibly diamictic

composition (Stewart & Stoker, 1990; Forwick & Vorren, 2011). AD5 could thus represent a diamict deposited either at or beneath the glacier grounding line as glacial till, or from increased iceberg-rainout. Our sedimentary data indicate that the diamict is more likely related to deposition from glacimarine processes (see LD1, section 4.3 below), but based on the partly distorted signal on the sub-bottom profiler data and a limited penetration depth of the cores into AD5, a clear distinction cannot be made.

4.3. Lithological data

4.3.1. Lithofacies

From the sedimentary record preserved in the vibrocores we define five lithofacies in Disko Bay (LD1–LD5), and one in the Vaigat (LV1). The correlation between lithology and sub-bottom profiler data is shown in Figure 5. Physical properties of the lithofacies and their stratigraphic distribution within the cores are displayed in Figures 6 and 8, while examples of the x-radiographs for each facies are shown in Figure 7.

LD1 is a dense ($2\text{--}3\text{ g cm}^{-3}$), matrix-supported diamict with a predominantly sandy matrix, and a majority of sub-angular to sub-rounded clasts (Fig. 7). Based on differences in shear strength and sediment structure, we distinguish LD1a and LD1b. LD1a shows some contortions on x-radiographs (Fig. 7), has a shear strength of up to 40 kPa, and only occurs in VC08 (Fig. 6). LD1b is massive, has a shear strength of up to 70 kPa, and only occurs in VC03 and VC04. The water content and the proportion of clay and silt in both lithofacies of LD1 are low with values around 20% and 40%, respectively (Fig. 6). Around 30-40% of the grains are $>250\text{ }\mu\text{m}$, with generally 5–10 clasts $>2\text{ mm}$ occurring per 2 cm-window. The MS is around $100 \times 10^{-5}\text{ SI}$ on average and shows distinct peaks. LD1b is part of AD5 (Fig. 5), but strong bedrock reflection hyperbolae on the TOPAS signal around the core site of VC08 prevented a direct correlation between LD1a and its acoustic counterpart.

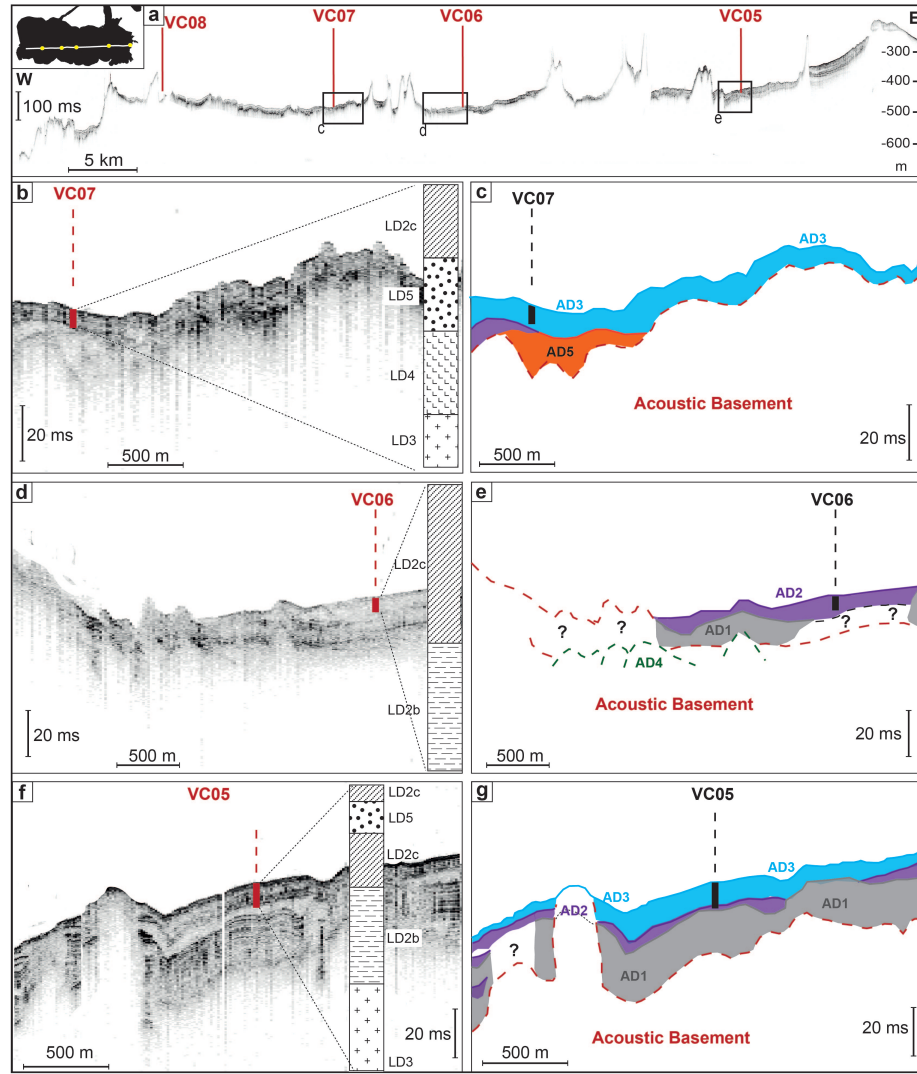


Figure 5: a) TOPAS profile across the core locations. The black polygon indicates location and extent of the profile. b), d), f) TOPAS lines across VC07, VC06, and VC05, respectively, with c), e), g) showing the according acoustic facies interpretation with respect to core penetration.

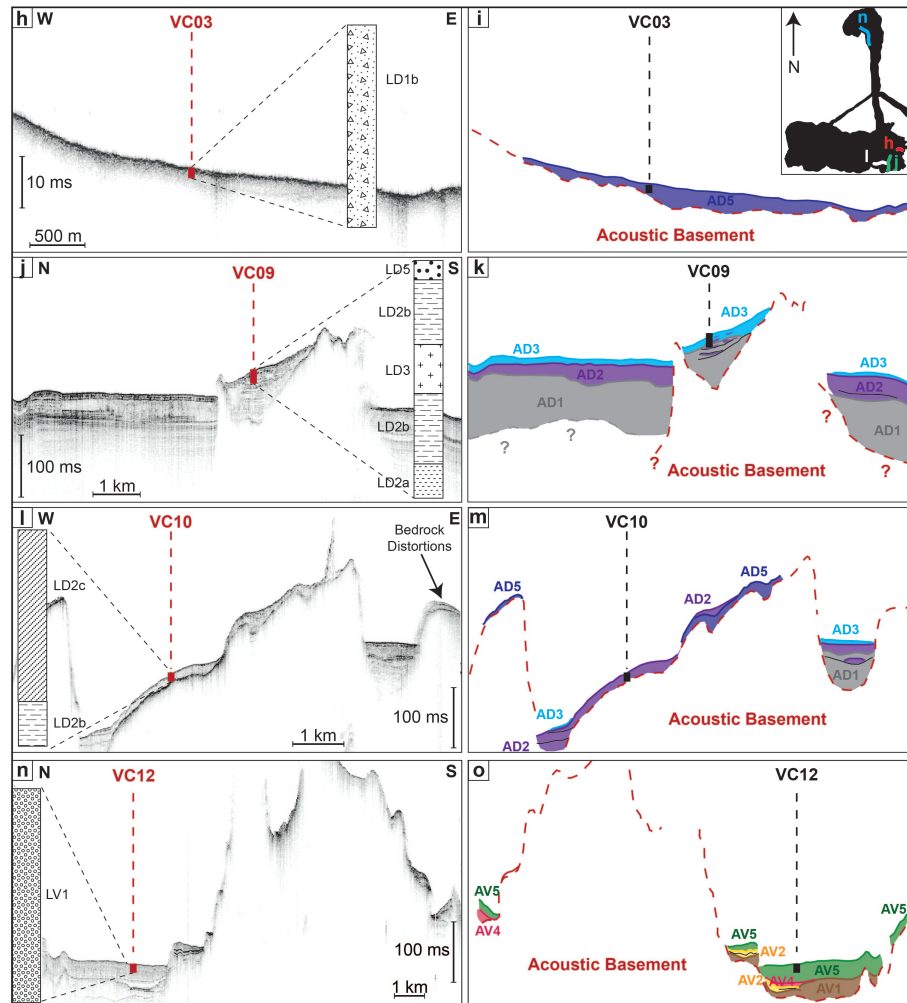


Figure 5 (cont.): TOPAS lines across and acoustic facies interpretation at the core sites of h), i) VC03, j), k) VC09, l), m) VC10, and n), o) VC12 from the Vaigat Strait. Black polygon in the top right-hand corner of i) shows the location of the TOPAS lines with respect to the bathymetry.

337 Lithofacies LD2 contains mud with highly variable amounts of clasts and is
 338 present in all cores from Disko Bay (Fig. 6). Clasts are pebble- to gravel-sized,
 339 mainly sub-angular to sub-rounded and of predominantly granitic composition,
 340 presumably sourced from the Precambrian basement (Fig. 7; cf. Chalmers et al.,
 341 1999; Larsen & Pulvertaft, 2000; Weidick & Bennike, 2007). The matrix is

342 composed of clay and silt and varies in colour between (dark) greenish grey
 343 (Munsell colour code: GLEY 1 4/10Y to 5/10Y) and greenish grey (GLEY
 344 1 5/10Y to 6/10Y) or dark to olive grey (5Y 4/1 to 4/2). The muds have
 345 a density of 1–2 g cm⁻³ and a shear strength between 2 and 10 kPa, which
 346 slightly increases down-core (Fig. 6). LD2 has a water content between 30
 347 and 60% (Fig. 6) and standing water was observed on localised areas of the
 348 sediment surface. The mud fraction generally exceeds 90% but can drop to 80%
 349 where clasts are abundant (Fig. 6). Clast concentrations are up to 25 clasts
 350 per 2 cm-window. The facies has a highly variable MS between ~15 and 150 x
 351 10⁻⁵ SI (Fig. 6). We distinguish subfacies LD2a, LD2b, and LD2c. In LD2a,
 352 which only occurs at the bottom of VC09, the mud appears diffusely stratified
 353 with some pebble-sized clasts and occasional bioturbation burrows at the top
 354 of the facies (Figs. 6, 7). LD2b contains internally massive mud in stratified
 355 sequences, with strata between ~4 and 15 cm thick. The strata have generally
 356 sharp contacts in the lower parts of LD2b, and more diffuse boundaries, partly
 357 promoted by bioturbation, in the upper parts (Figs. 6, 7). LD2c contains
 358 massive, occasionally bioturbated mud (Fig. 7). LD2 correlates with acoustic
 359 facies AD2 and AD3 (Fig. 5).

360 Lithofacies LD3 is composed of massive and partly contorted mud, inter-
 361 persed with massive fine sand-rich units (Fig. 6). These units are occasional
 362 to frequent, mostly inclined, and occur as mm-thick laminae or cm-thick layers
 363 with sometimes sharp, but mostly diffuse lower boundaries (Fig. 7). In places
 364 the sandy beds are heavily contorted (Fig. 7). The overall density of LD3 is
 365 around 1.6 g cm⁻³ with minor variations, whereas the shear strength is highly
 366 variable (0.5 and 12 kPa; Fig. 6). Water content is around 20–30%, and grain
 367 size distribution varies according to the sub-sampled lithology, with >95% clay
 368 and silt in the matrix, and ~80% clay and silt in the sandy layers (Fig. 6). IRD
 369 grains >2 mm are rare. The MS is ~100–120 x 10⁻⁵ SI with few localised and
 370 distinct peaks. The facies occurs in either the basal or middle parts of VC05,
 371 VC07, and VC09 (Fig. 6). LD3 forms part of the acoustic facies AD2 (Fig. 5).

372 Lithofacies LD4 contains diffusely laminated mud interbedded with diamictic

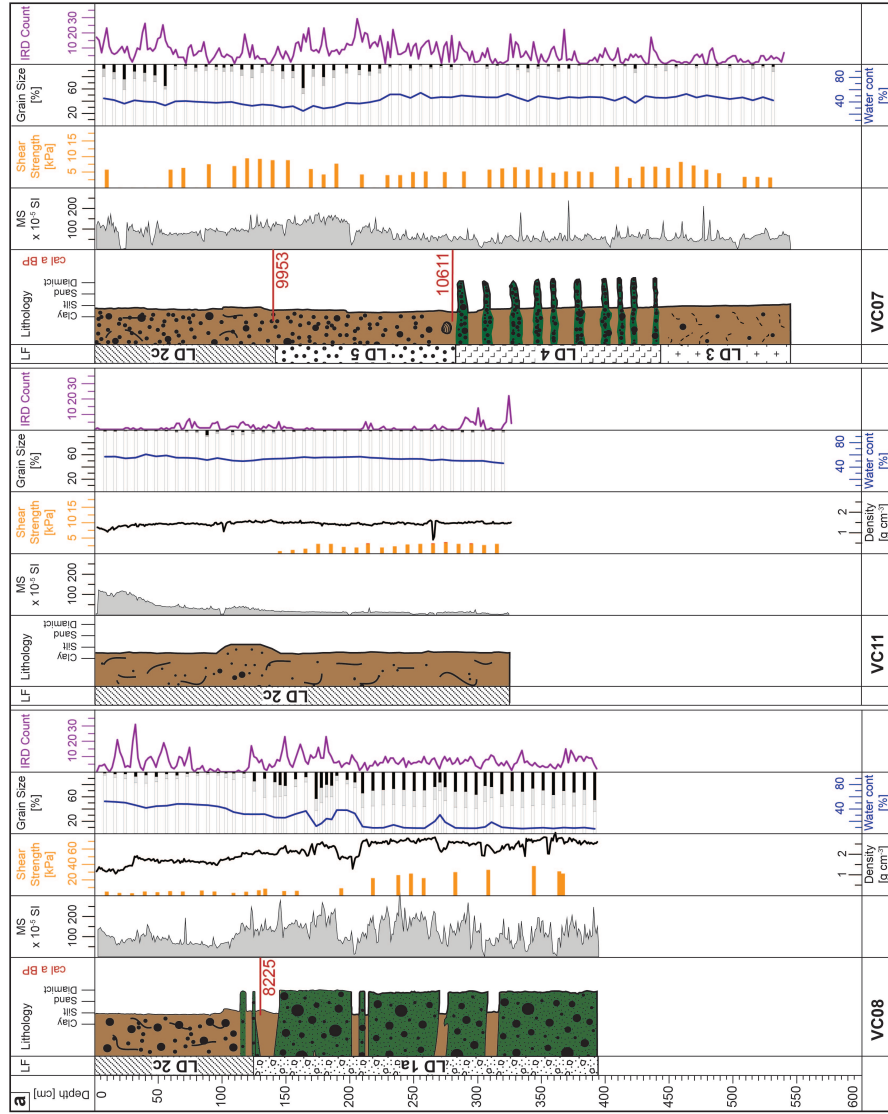


Figure 6: a) Lithological and lithofacies logs with physical properties of vibrocores VC08, VC11, and VC07 from Disko Bay (west to east). MS = magnetic susceptibility. Note the different scales for the shear strength. Grain-size distribution results were grouped thus: white bars = grains <63 μm , grey bars = 63–250 μm , black bars = grains > 250 μm .

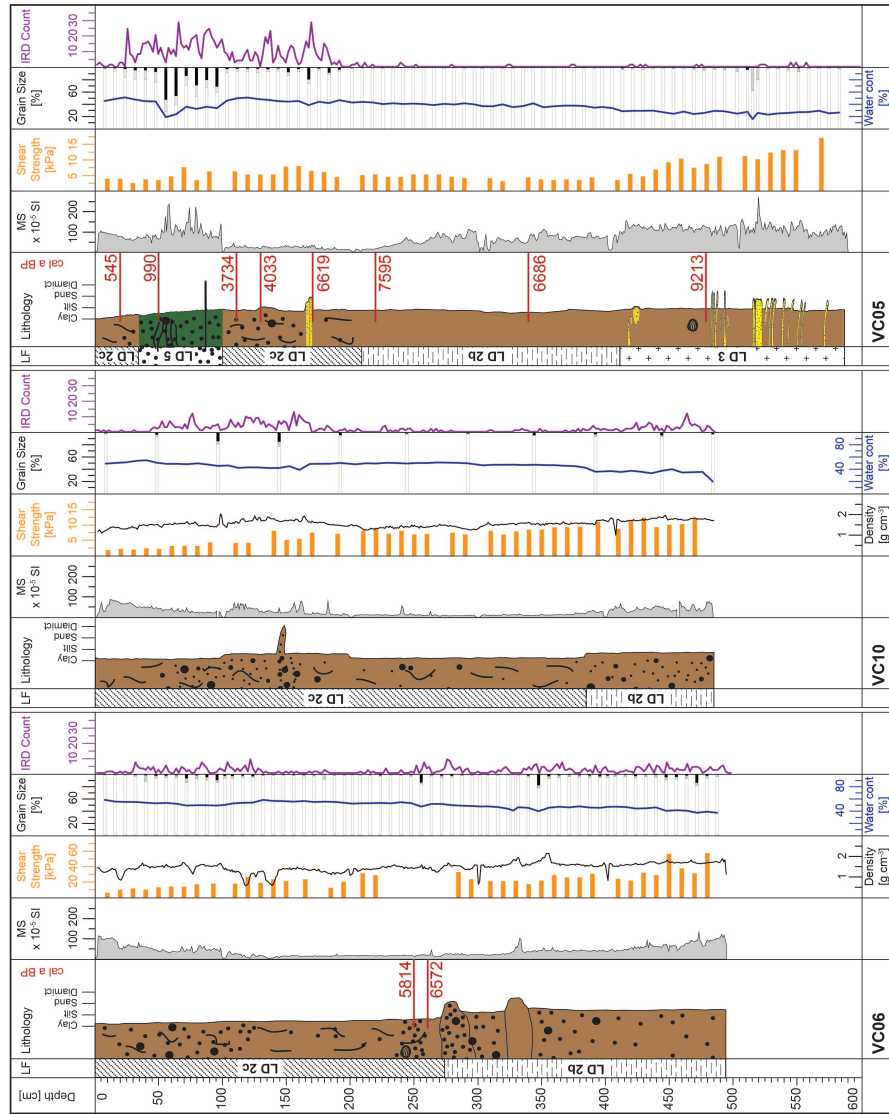


Figure 6 (cont.): Lithological and lithofacies logs with physical properties of vibrocores VC06, VC10, and VC05 from Disko Bay (west to east). MS = magnetic susceptibility. Grain-size distribution: white bars = grains <63 μm , grey bars = 63–250 μm , black bars = grains > 250 μm .

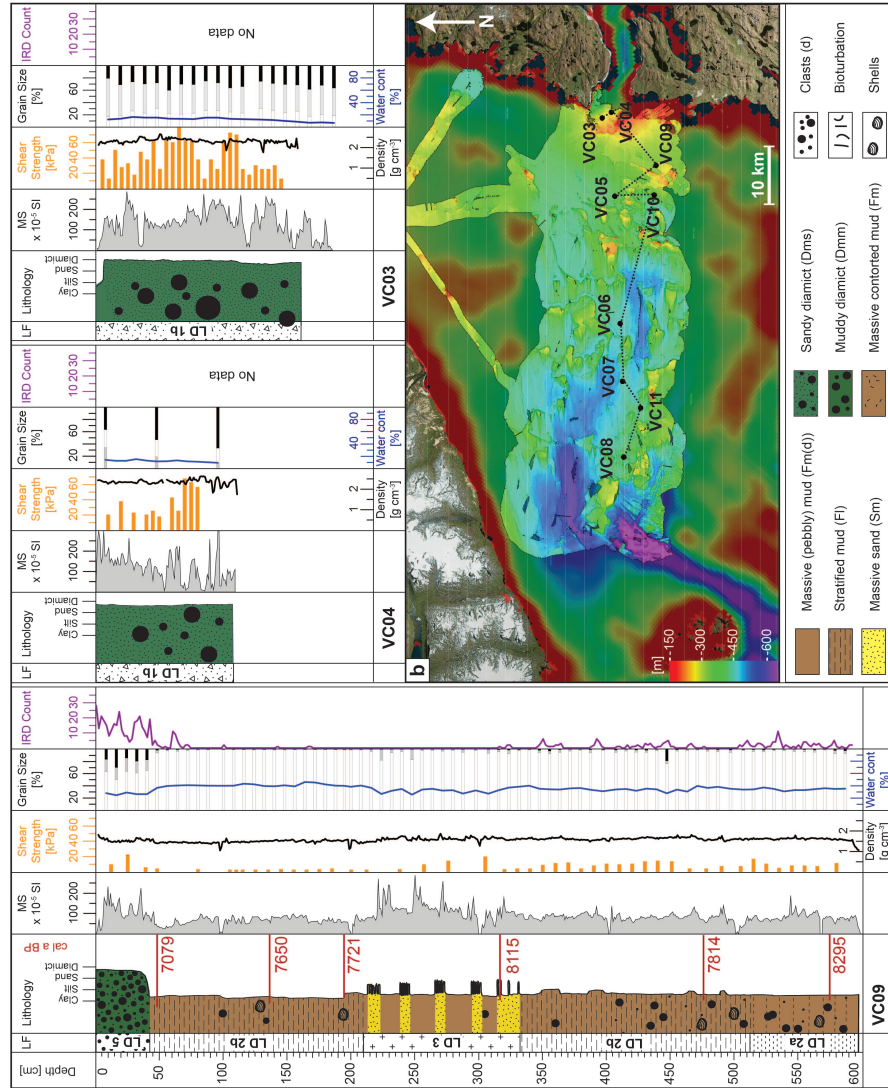


Figure 6 (cont.): Lithological and lithofacies logs with physical properties of vibrocores VC09, VC04, and VC03 from Disko Bay (west to east). MS = magnetic susceptibility. b) Core locations. The black stippled line indicates the order in which cores are shown. Grain-size distribution: white bars = grains <63 μm , grey bars = 63–250 μm , black bars = grains > 250 μm .

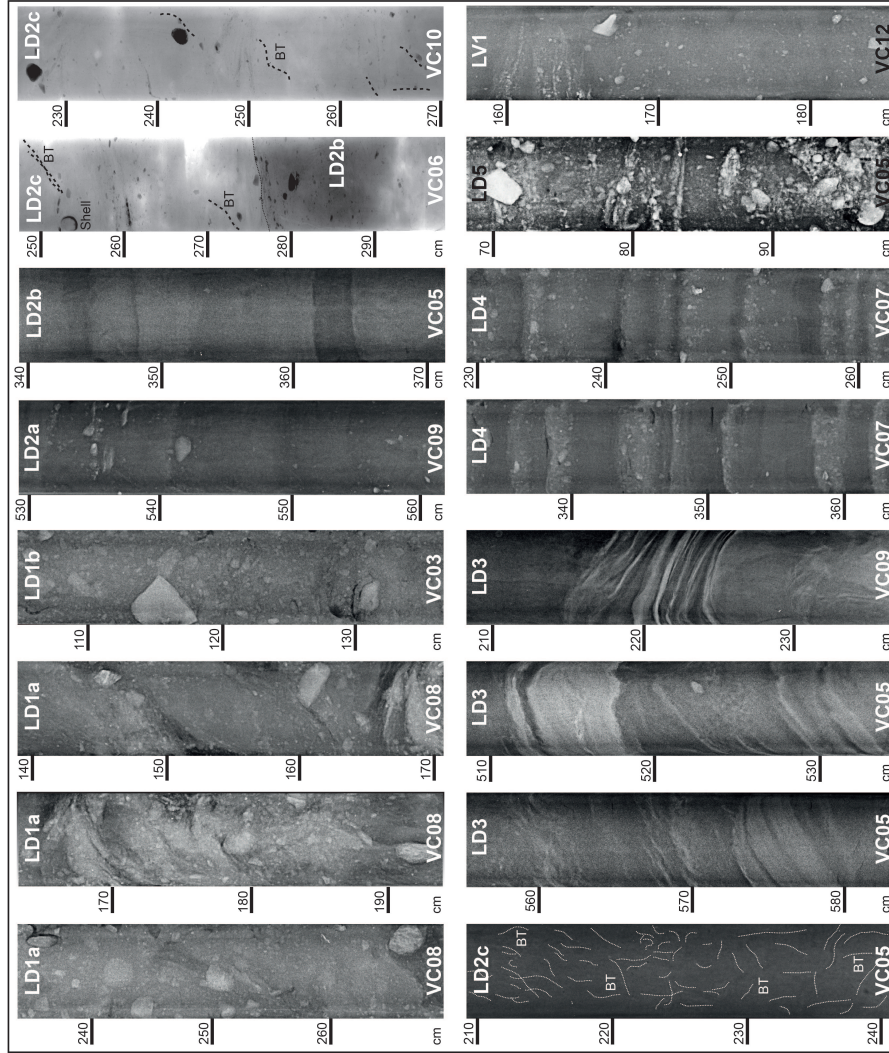


Figure 7: Examples of the x-radiographs from the six lithofacies LD1–LD5, and LV1 from different vibrocores (VC) and sediment depths in Disko Bay and the Vaigat Strait. Generally denser parts appear lighter, except for in the two x-radiographs from VC10 and VC06, where colours are reversed.

373 layers (Figs. 6, 7). The diamict layers are up to 4 cm thick, contain fines, sand,
 374 and pebbles, and normally have a very gradual and diffuse lower boundary, but
 375 a sharper upper contact, which is especially pronounced in the bottom parts of
 376 LD4 (Fig. 7). The shear strength is around 8 kPa, the water content $\sim 40\%$, and
 377 $>90\%$ of the sediment are $<63\mu\text{m}$. The increased clast content in the diamictic
 378 layers is reflected in a variable IRD count of up to 21 clasts per 2 cm (Fig.
 379 6). Minor oscillations in MS, which is generally $\sim 50 \times 10^{-5}$, are interrupted by
 380 several pronounced peaks to values between 125 and 225×10^{-5} SI (Fig. 6).
 381 LD4 is present only in VC07 (Fig. 6).

382 Lithofacies LD5 is a matrix-supported diamict, with a mud-dominated ma-
 383 trix (rather than sand-dominated as for LD1) and abundant angular to sub-
 384 angular clasts of variable diameter (Fig. 7). Smaller pebbles are sometimes
 385 concentrated in dense, cm-thick beds, which have sharp contacts with the sur-
 386 rounding sediments. The density is around 1.5 g cm^{-3} , the shear strength
 387 between 5 and 10 kPa, and the water content between 10 and 40%. LD5 has a
 388 variable proportion (20–80%) of fines ($<63 \mu\text{m}$) and a high IRD count of up to
 389 25 clasts per 2 cm-window. A spiky appearance with numerous well-pronounced
 390 peaks defines the MS, which, in most cases, exceeds 80×10^{-5} SI.

391 The cores from the Vaigat contain a single lithofacies, LV1, which comprises
 392 massive, dark grey to dark olive grey mud (5Y 4/1 to 3/2) with a moderate
 393 clast abundance (Figs. 7, 8). It is similar to LD2a from the cores in Disko
 394 Bay, but the matrix contains low amounts of sand and there is no evidence of
 395 bioturbation. Internal sedimentary structures are absent, with the exception
 396 of occasional laminae of coarser grains (medium sand to pebble-sized; Fig. 7).
 397 The density is around 1.5 g cm^{-3} and the shear strength 4–12 kPa, with a
 398 slight increasing trend down-facies (Fig. 8). LV1 has a water content of $\sim 50\%$
 399 and $>95\%$ fines and occurs in all three cores from the Vaigat (Fig. 8). The
 400 IRD count is highly variable with 0–22 clasts per 2 cm-window, and the MS is
 401 generally around $200\text{--}250 \times 10^{-5}$ SI with relatively minor variations (Fig. 8).
 402 LV1 forms part of acoustic facies AV5 (Fig. 5).

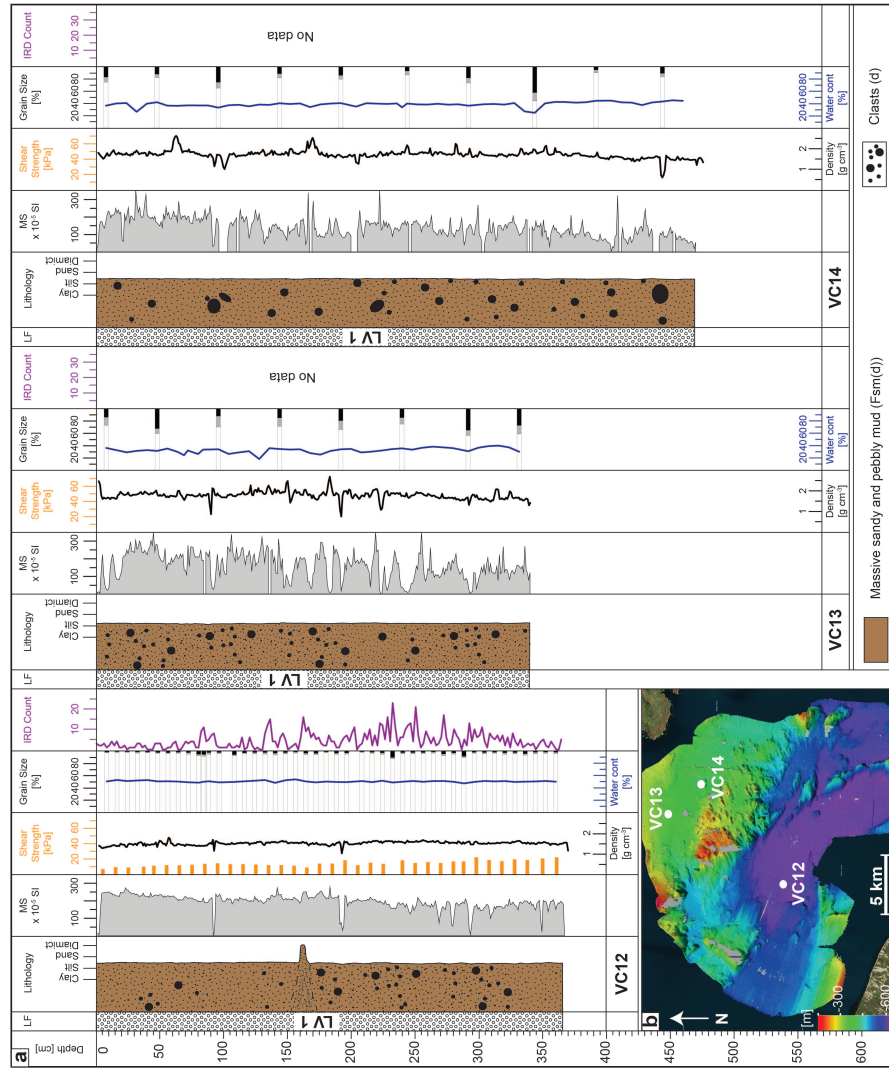


Figure 8: Lithological and lithofacies logs of the three vibrocores from the Vaigat Strait and their physical properties. MS = magnetic susceptibility. Grain-size distribution: white bars = grains <63 μm , grey bars = 63–250 μm , black bars = grains > 250 μm .

4.3.2. Radiocarbon dates and sediment accumulation rates

Table 2: Radiocarbon dates and calibrated ages used in this study. Unless otherwise specified all bivalves were intact and did not show evidence of being re-worked.

Core ID	Depth [cm]	Lab Code	Sample	Reported age [^{14}C a BP]	Mean probability age [cal a BP]	2σ [cal a BP]
VC05	20–21	AA-90391	Seaweed	1079 ± 78	545	418–668
VC05	50–51	AA-90392	Seaweed	1575 ± 88	990	778–1190
VC05	112–113	Beta-434927	Seaweed	3930 ± 30	3734	3612–3845
VC05	130–131	AA-90393	Seaweed	4159 ± 50	4033	3872–4210
VC05	170–171	AA-90394	Seaweed	6322 ± 60	6619	6451–6776
VC05	220–221	AA-90395	Seaweed	7250 ± 380	7595	6788–8355
VC05	340–341	AA-90396	Seaweed	6370 ± 180	6686	6294–7119
VC05	478–479	AA-90396	Paired Bivalve	8710 ± 50	9213	9045–9383
VC06	250	Beta-434928	Single Bivalve	5580 ± 30	5814	5711–5905
VC06	272–273	Beta-434929	Seaweed	6280 ± 30	6572	6456–6672
VC07	140–142	Beta-434930	Foraminifera	9300 ± 30	9953	9772–10118
VC07	280–281	Beta-434931	Paired Bivalve	9850 ± 30	10611	10503–10723
VC08	130–132	Beta-434932	Foraminifera	7890 ± 30	8225	8130–8326
VC09	48	Beta-434933	Seaweed	6700 ± 30	7079	6955–7184
VC09	137	Beta-434934	Paired Bivalve	7330 ± 30	7650	7570–7743
VC09	195	Beta-434935	Paired Bivalve	7400 ± 30	7721	7631–7821
VC09	316–318	Beta-434936	Foraminifera	7800 ± 30	8115	8001–8214
VC09	476	Beta-265208	Paired Bivalve	7490 ± 50	7814	7684–7928
VC09	575	Beta-265209	Paired Bivalve	7970 ± 50	8295	8171–8394

Radiocarbon dates presented here are rounded to the closest 100 years and are shown in detail in Table 2 and Figure 9. AMS measurements were carried out on paired or whole bivalve shells (Table 2) or mixed benthic foraminifera taken from seemingly undisturbed sediment sections to ensure accuracy of the radiocarbon dates. The exception is the date at 316 cm in VC09, where the frequent sandy deposits most likely represent turbidites and the obtained date may hence derive from reworked material (see section 4.3.3 below). Other AMS measurements were carried out on seaweed taken from near the centre of the split cores to reduce the risk of material having been dragged down-core during the coring process. Notwithstanding this, the age reversal and the large error margin for the sample from 340 cm in VC05 indicate that the seaweed at this depth

was not in-situ. Sediment accumulation rates (SARs; Fig. 9) were calculated from the mean radiocarbon ages and assume constant accumulation between each date.

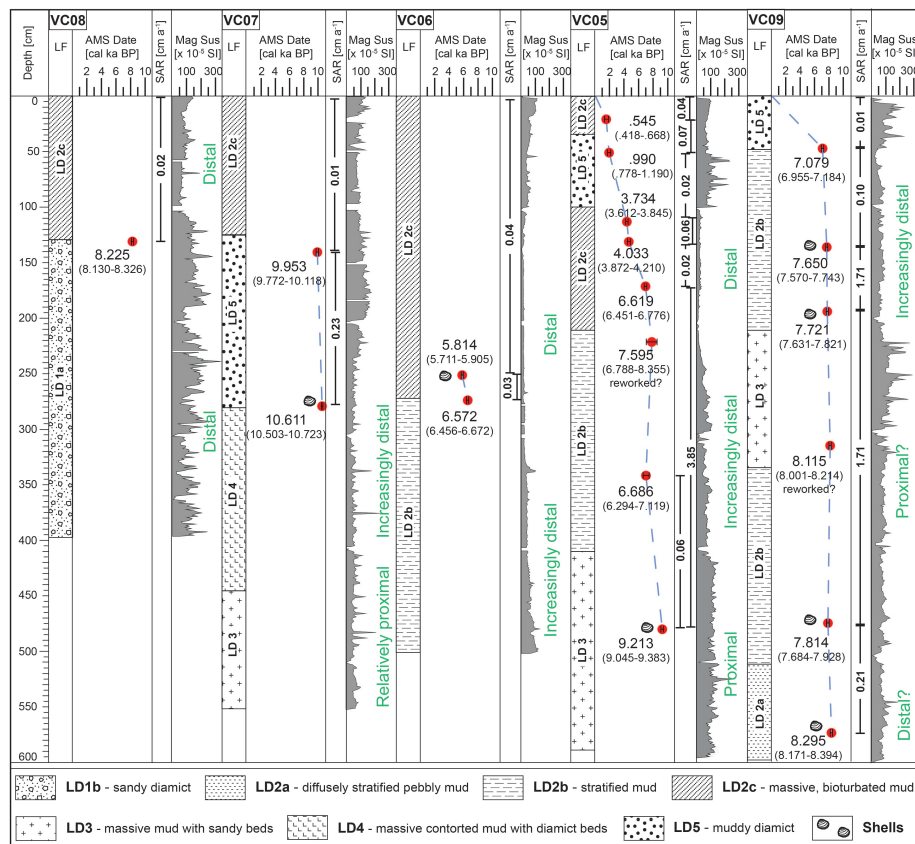


Figure 9: Lithofacies logs with radiocarbon dates (error bars indicate 2σ -range) and calculated SARs from a core transect across Disko Bay from west to east (see Fig. 6b for core locations). Note that core signatures are only used as a means to visually differentiate between lithofacies, not to describe the actual lithology. For details on the lithological composition of the cores and lithofacies see also Fig. 6 and section 4.3. Together with the MS these values give an idea about the glacial marine environment the lithofacies were deposited in.

Radiocarbon dating suggests that LD1a at the base of VC08 was deposited prior to ~ 8.2 cal ka BP and that the overlying LD2c accumulated at an average rate of 0.02 cm a^{-1} (Fig. 9). Basal sediments in VC07 contain LD3 and LD4

421 and are the oldest recovered from Disko Bay, deposited before 10.6 cal ka BP.
 422 The top of the overlying LD5 dates to ~ 9.9 cal ka BP, implying a SAR of 0.23
 423 cm a^{-1} for this facies (Fig. 9). The same date suggests that the topmost facies
 424 in VC07, LD2c, was deposited afterwards at a SAR of 0.01 cm a^{-1} . Two dates
 425 from VC06 indicate that its bottom part (LD2b) is older than ~ 6.5 cal ka BP,
 426 and that the overlying LD2c accumulated at a rate of $0.03\text{--}0.04 \text{ cm a}^{-1}$ (Fig. 9).
 427 LD3 at the bottom of VC05 was deposited around 9.2 cal ka BP. Assuming that
 428 the date of 7.6 cal ka BP at the top of the overlying LD2b is indeed unreliable,
 429 LD2b accumulated from before 6.7 to just before 6.6 cal ka BP at a rate of
 430 $\sim 0.05 \text{ cm a}^{-1}$ (Fig. 9). The overlying LD2c was deposited until today, at a rate
 431 of $0.02\text{--}0.06 \text{ cm a}^{-1}$, but deposition was interrupted from ~ 3.7 until after 0.99
 432 cal ka BP, during which time LD5 formed (Fig. 9). The bottom half of VC09
 433 (LD2a, LD2b and LD3) was deposited between 8.3 and 7.7 cal ka BP at an SAR
 434 of $0.21\text{--}1.71 \text{ cm a}^{-1}$ (Fig. 9). In the top half of the core LD2b accumulated
 435 between 7.7 and 7.0 cal ka BP at decreasing rates between 1.72 and 0.10 cm a^{-1} ,
 436 and is overlain by LD5, deposited since 7.0 cal ka BP at a very low SAR
 437 of 0.01 cm a^{-1} (Fig. 9).

438 **4.3.3. Interpretation**

439 **LD1 – sandy diamict**

440 Although x-radiographs of LD1 are similar to glacial till documented from outer
 441 Disko Bay and the continental shelf in the Disko and Uummannaq Troughs
 442 (Ó Cofaigh et al., 2013; Dowdeswell et al., 2014; Hogan et al., 2016; Sheldon
 443 et al., 2016) and the shear strength of $\sim 40\text{--}80 \text{ kPa}$ exceeds that of subglacial
 444 tills from Antarctica (cf. Ó Cofaigh et al., 2005), the absence of planar struc-
 445 tures or indications of clast alignment in both, LD1a and LD1b appear to be at
 446 odds with an interpretation as glacial till. Furthermore, in the case of LD1a, we
 447 would expect glacial till to be significantly older than the given age of 8.2 cal ka
 448 BP, considering that the bay was probably ice-free around 10 cal ka BP (Lloyd
 449 et al., 2005; Young et al., 2011a). In the case of LD1b, we would expect any

sequence of glacial till to be covered by a succession of Holocene sediments, as
till could only have formed at the core sites when the ice margin last paused at
Isfjeldsbanken, around 10 ka BP. We therefore favour an interpretation of LD1a
as a mass-flow deposit and of LD1b as post-glacial glacimarine sediments, where
deposition occurred from meltwater and/or the water column and melting ice-
bergs (e.g. Elverhøi et al., 1980, 1983; Gilbert, 1983; Hogan et al., 2016). The
partly contorted appearance of LD1a in VC08 and the occurrence of mud strata
within the facies indicate sediment reworking and suggest that LD1a was formed
as a gravity-flow deposit reworking glacimarine mud and IRD (cf. e.g. Kuenen,
1948; Shanmugam et al., 1994; Forsberg et al., 1999). This seems reasonable,
as the core site of VC08 is located in a submarine basin between two bedrock
highs, the steep slopes of which could have promoted repeated mass-flows. In-
deed, the sub-bottom profiler data show abundant mass-flow deposits in Disko
Bay, supporting an interpretation of LD1a as a gravity-flow deposit. This inter-
pretation is also consistent with the low shear strength and the comparatively
young age of LD1a. An interpretation of LD1b as post-glacial IRD-rich sediment
is based on the fact that VC03 and VC04, containing LD1b, were taken from
the top of Isfjeldsbanken, where the high accumulation of IRD is likely, because
(1) all icebergs calved off Jakobshavn Isbræ pass through Isfjorden into Disko
Bay, and (2) the sill traps larger icebergs inside the fjord due to the shallow
water depth (Echelmeyer et al., 1991; Hogan et al., 2011). These bergs need
to melt considerably before passing into Disko Bay (Echelmeyer et al., 1991;
Hogan et al., 2011), and thus probably deposit the majority of their debris at or
near Isfjeldsbanken. Although no distinct iceberg ploughmarks were observed
on the swath-bathymetric data, the relatively high shear strength in LD1b indi-
cates that icebergs may have grounded at the core sites, scouring, compacting
and homogenising the deposited sediments. Indeed, LD1b is macroscopically
similar to diamicts in East Greenland, which have been interpreted as "iceberg
turbates" (Vorren et al., 1983; Marienfeld, 1992; Dowdeswell et al., 1994; Linch
& Dowdeswell, 2016).

480 **LD2 – massive to stratified mud**

481 LD2 is interpreted as glacimarine mud settling from suspension in glacial melt-
482 water plumes and the water column (cf. e.g. Gilbert, 1983; Elverhøi et al., 1983;
483 Dowdeswell et al., 1998; Ashley & Smith, 2000). Clasts record the deposition
484 of IRD from icebergs and/or sea ice. The low shear strength and high water
485 content of this facies support this interpretation.

486 Stratification in LD2a and LD2b is likely related to changes in depositional
487 environment or sediment source. These are controlled by a number of factors,
488 including e.g. regional warming and melting of the ice sheet, variability in the
489 position of meltwater efflux, inter-annual variations in meltwater flux due to sea-
490 sonal or tidal controls or floods from glacial lakes, or by ice-margin fluctuations
491 controlling the proximity of the core site to the ice margin. The diffuse nature
492 of the stratification in LD2a is thought to reflect a relatively ice-distal environ-
493 ment, where such changes have a smaller impact on the sedimentary record.
494 An upward increase in bioturbation and the low SAR of 0.2 cm a^{-1} support
495 an interpretation of LD2a as glacimarine mud deposited under (increasingly)
496 distal conditions. Conversely, the distinct stratification in LD2b is thought to
497 be indicative of a more glacier-proximal depositional environment. Ice-proximal
498 conditions are supported by a high concentration of meltwater-derived fines, also
499 reflected in SARs of up to 1.7 cm a^{-1} , an increased MS suggesting a predom-
500 inantly terrestrial input (Steenfelt et al., 1990; Robinson, 1993; Møller et al.,
501 2006; Seidenkrantz et al., 2013), and the lack of IRD, as under these conditions
502 the input from icebergs is often masked by the fast accumulation of meltwater-
503 derived fines (e.g. Elverhøi et al., 1980; Dowdeswell & Dowdeswell, 1989; Cowan
504 et al., 1997; Gilbert et al., 2002). Radiocarbon dates show that LD2a was de-
505 posited some time after 8.2 cal ka BP, and accordingly the change from more
506 distal to more proximal conditions between LD2a and LD2b in VC09 possibly
507 reflects a re-advance of Jakobshavn Isbræ in response to the 8.2 ka cooling event
508 (Weidick & Bennike, 2007; Young et al., 2011a, 2013). LD2b was deposited prior
509 to 7.6 cal ka BP in VC05 and between ~ 7.8 and 7.1 cal ka BP in VC09 (inter-
510 rupted by deposition of LD3 just before 7.7 cal ka BP, Fig. 9). During this time

511 Jakobshavn Isbræ was at Isfjeldsbanken and began retreating into Isfjorden,
 512 which could have resulted in minor oscillations in the position of the ice margin
 513 causing the stratification in LD2b. However, deposition of LD2b also coincides
 514 with a period of strongly increased meltwater discharge, presumably caused by
 515 extensive thinning of the ice sheet prior to 8.3 ka BP (Rinterknecht et al., 2009;
 516 Seidenkrantz et al., 2013). Stratification would then have been imparted by
 517 inter-seasonal variations in meltwater flux to the core site, as the large variability
 518 in strata thickness within LD2b is at odds with the rhythmic stratification
 519 usually observed in seasonally-controlled meltwater deposits (cf. e.g. Domack,
 520 1984; Mackiewicz et al., 1984; Ó Cofaigh & Dowdeswell, 2001). Although both
 521 scenarios are possible for the deposition of LD2b, we favour the former possibility
 522 and suggest that LD2b was deposited in a glacier-proximal environment
 523 where stratification was caused by minor ice front oscillations. This is based on
 524 several reasons, including that: (1) the occurrence of presumably distal glaci-
 525 marine mud (LD2a) at the bottom of VC09 around 8.2 cal ka BP is strange given
 526 that enhanced meltwater release started as early as 8.6 ka BP in some regions,
 527 (2) lithological evidence from Disko Bay showed that the meltwater event had
 528 ceased by 7.7–7.5 ka BP (Seidenkrantz et al., 2013), yet deposition of LD2b and
 529 hence a strong meltwater signal prevailed until 7.0 cal ka BP in VC09, and even
 530 until 6.6 cal ka BP in VC06, and (3) if the meltwater event was responsible
 531 for the deposition of LD2b, the deposition of LD3 in between two packages of
 532 LD2b in VC09 (Fig. 9) would be difficult to understand (see also below). Furthermore,
 533 the traces of bioturbation at the top of LD2b and the declining MS
 534 and SARs up-facies are more easily accounted for by gradual ice margin retreat
 535 rather than a drastic reduction in meltwater flux. Ongoing retreat could also
 536 have led to an increasing amount of sediment being trapped behind the sill in
 537 Isfjorden, thus causing additional variability in the depositional environment at
 538 the core sites.

539 Based on its massive structure and the presence of bioturbation burrows
 540 suggesting favourable living conditions for some benthic organisms, LD2c is
 541 interpreted as ice-distal glaci-marine mud. This is in accordance with the ra-

542 diocarbon dates, which provide evidence for deposition of LD2c after ~ 6.7 cal
 543 ka BP in VC05, during which time ice was retreating through Isfjorden (e.g.
 544 Lloyd et al., 2005; Weidick & Bennike, 2007; Hogan et al., 2011). The switch
 545 from ice-proximal (LD2b) to ice-distal conditions (LD2c) between ~ 6.7 and 7.6
 546 cal ka BP in VC05 and around 7.1 cal ka BP in VC05 and VC09, respectively,
 547 show that by this time Jakobshavn Isbræ had retreated so far into Isfjorden,
 548 that meltwater sedimentation was no longer the dominant process at the core
 549 sites. The stratigraphic position of LD2c at the top of most cores from Disko
 550 Bay, the low SAR of $\sim 0.2 \text{ cm a}^{-1}$, and the low MS indicating a predominantly
 551 hemipelagic origin (cf. Steenfelt et al., 1990; Møller et al., 2006; Seidenkrantz
 552 et al., 2013) support this interpretation (e.g. Syvitski & Murray, 1981; Gilbert,
 553 1982; Boulton, 1990; Sexton et al., 1992). Indeed, the ice margin is thought to
 554 have retreated behind its present position during the mid-Holocene, where it
 555 remained until ~ 2.2 ka BP (Weidick & Bennike, 2007). However, the gradual
 556 increase in MS in the top 50 cm of LD2c in most cores from Disko Bay, and the
 557 simultaneous increase in SARs in VC05 and VC06 (Fig. 9) suggest a terrestrial
 558 origin for the youngest sediments in Disko Bay and a relatively higher availabil-
 559 ity of meltwater-derived fines. This could be related to the westward re-advance
 560 of the ice margin after 2.2 ka BP and/or the recent increase of thinning and sub-
 561 glacial melting of Jakobshavn Isbræ and adjacent GIS outlets (Holland et al.,
 562 2008; Rignot et al., 2010).

563 **LD3 – stratified mud with sand laminae**

564 The stratified fine-grained mud in lithofacies LD3 is interpreted as glacima-
 565 rine mud from meltwater, with the sand laminae deposited from down-slope
 566 gravity flows, e.g. turbidity currents (e.g. Gilbert, 1982; Elverhøi et al., 1983;
 567 Mackiewicz et al., 1984; Sexton et al., 1992; Ó Cofaigh & Dowdeswell, 2001).
 568 Where the sand and mud layers appear contorted, it likely relates to gravita-
 569 tional slump events that acted to rework and redeposit the sediments down-slope
 570 (e.g. Kuenen, 1948; Shanmugam et al., 1994; Forsberg et al., 1999). Turbidites
 571 are often associated with proximal glacimarine conditions (e.g. Gilbert, 1982;

572 Gilbert et al., 1993), suggesting that during the deposition of LD3 the margin
 573 of Jakobshavn Isbræ was relatively close. Although the turbidites could simply
 574 be a product of the abundant mass-flows occurring in Disko Bay due to the
 575 very irregular topography, an ice-proximal origin for LD3 is also supported by
 576 the radiocarbon dates, which provide evidence that LD3 was deposited around
 577 9.2 cal ka BP in VC05, and between 7.8–7.7 ka BP in VC09 (the date of 8.1
 578 ka BP in VC09 is considered unreliable, see section 4.3.2), during which time
 579 the ice margin was at or near Isfjeldsbanken (Lloyd et al., 2005; Long et al.,
 580 2006; Weidick & Bennike, 2007; Kelley et al., 2013). An ice-proximal origin
 581 for LD3 is further supported by the overall constant, relatively high MS, the
 582 close stratigraphic relationship of LD3 to the proximal sediments of LD2b (Fig.
 583 6), and the lack of IRD, as large amounts of fresh glacial meltwater may have
 584 promoted debris retention in icebergs, or high SARs of mud may have swamped
 585 the input of IRD.

586 **LD4 – stratified mud and diamict**

587 The inter-stratified muds and diamicts of lithofacies LD4 are similar to deposits
 588 described from the continental shelf west of Disko Bay, which were linked to
 589 variations in meltwater-derived sediment flux, likely related to seasonal cycles
 590 (Hogan et al., 2016). As also argued by Hogan et al. (2016), such deposits can
 591 form in two ways: (1) the fine-grained layers form during summer, when IRD
 592 flux to the core sites is overwhelmed by increased influx of meltwater-derived
 593 fines. In winter, sedimentation from icebergs dominates over the reduced influx
 594 of fine-grained mud associated with lower melt rates, and the diamictic layers
 595 form (cf. Cowan et al., 1997; Hogan et al., 2016). (2) The fine-grained sediments
 596 are deposited during winter, when shore-fast sea ice traps icebergs and prevents
 597 the meltout and deposition of IRD, while subglacial meltwater delivers fine-
 598 grained mud to the core sites. In summer the sea ice breaks up and releases the
 599 icebergs, and their incorporated IRD melts out and forms the diamict layers
 600 (cf. Syvitski et al., 1996; Dowdeswell et al., 1998, 2000). The generally sharper
 601 upper boundaries of the diamict units in the lower parts of LD4 indicate an

abrupt increase in meltwater discharge at the start of the summer season and/or
 increased sediment concentrations in the meltwater plumes (cf. Ó Cofaigh &
 Dowdeswell, 2001; Knudsen et al., 2007) and we thus suggest that the mud layers
 in LD4 were more likely deposited from a meltwater plume (cf. Hogan et al.,
 2016). The change to noticeably more diffuse boundaries towards the upper
 part of LD4 likely represents a transition from a more proximal glacimarine
 environment lower in LD4 to more distal conditions higher up in the facies (cf.
 Hogan et al., 2016). A radiocarbon date from LD4 indicates that this facies
 was deposited prior to 10.6 ka BP (Fig. 9), which supports this interpretation
 as the ice margin was probably slightly east of Isfjeldsbanken during this time
 (Lloyd et al., 2005). The date further provides an important constraint for the
 retreat of Jakobshavn Isbræ through Disko Bay (see section 5.2).

LD5 – pebbly mud

The muddy matrix in facies LD5 is interpreted as the product of hemipelagic
 or distal glacimarine suspension settling, with the predominantly angular clasts
 likely deposited from icebergs (cf. Dowdeswell & Dowdeswell, 1989). The high
 clast abundance points to (1) increased iceberg calving rates, (2) concentrated
 dumping events (i.e. overturning icebergs), (3) increased iceberg melt, or (4)
 a decreasing accumulation of fine-grained sediments from a retreating ice mar-
 gin emphasising the IRD input (e.g. Elverhøi et al., 1980, 1983; Dowdeswell &
 Dowdeswell, 1989; Moros et al., 2002; Andresen et al., 2010; Seidenkrantz et al.,
 2013). The occurrence of LD5 in VC07 and VC09 does not seem to be linked to a
 specific climatic warming event, which suggests iceberg dumping to be the most
 likely cause of formation. Conversely, the deposition of LD5 dates to ~ 3.7 ka BP
 in VC05 during which time the Jakobshavn ice margin was located east of its
 present position (Weidick & Bennike, 2007). As some authors have suggested
 a late Holocene Thermal Maximum in Disko Bay and other West Greenland
 fjords lasting until at least ~ 3.5 ka BP (Moros et al., 2006; Møller et al., 2006;
 Seidenkrantz et al., 2007), the deposition of LD5 in this core could be linked to
 enhanced glacier and/or iceberg melting during the late stages of this climatic

amelioration. As argued by Long & Roberts (2003), Roberts & Long (2005), and Lloyd (2006), warmer conditions would have led to reduced coupling between the glacier and its bed, and thus reduced supply of meltwater-derived fines to the core sites. The latter could be reflected in the low SARs ($\sim 0.02 \text{ cm a}^{-1}$) of LD5. Incidentally, although not specifically mentioned in the literature, such brief periods of climatic warming, or, alternatively, strengthened inflow of warm WGC waters, seem plausible in a system as climatically complex as Disko Bay and could thus also account for the occurrence of LD5 in VC07 and VC09. Notwithstanding this, the high concentration of IRD in LD5 could also imply a re-advance of Jakobshavn Isbræ, possibly related to the onset of Neoglaciation, which would have caused the ice to be subjected to the increasing influence of warm Atlantic water, leading to enhanced calving rates. The high flux of IRD would then have outpaced the flux of meltwater, leading to the deposition of a clast-rich mud. A higher iceberg availability was one of the reasons suggested for the high amounts of IRD in the lithological record from the Vaigat Strait (Andresen et al., 2010).

LV1 – massive mud with clasts

Based on its similarity to LD2a and sediments from the same area investigated by Andresen et al. (2010) and McCarthy (2011), we interpret LV1 from the Vaigat Strait as glacial-marine mud deposited from meltwater plumes and/or the water column, and clasts settling from icebergs and sea ice drifting over the core sites. Large-scale re-working of the sediments down-slope into the Vaigat Strait probably led to the massive internal structure. The comparatively large number of clasts in LV1 can be explained by the fact, that, in addition to the icebergs from the local glaciers, icebergs calved from Jakobshavn Isbræ are also transported to the Vaigat Strait by the local ocean currents (Lloyd et al., 2005; Andresen et al., 2010; McCarthy, 2011).

659 5. Discussion

660 5.1. Landforms and sediment facies signature of Jakobshavn 661 Isbræ and adjacent fast-flowing GIS outlets

662 The submarine landform assemblage in Disko Bay and the Vaigat Strait includes
663 (1) streamlined bedrock ridges, (2) crag-and-tails, (3) submarine channels, and
664 (4) pockmarks. TOPAS profiles further indicate the abundance of (5) deposits
665 from gravitational mass-transport. The streamlined bedrock and crag-and-tails
666 record the flow of an extended Jakobshavn Isbræ and adjacent GIS outlets
667 across the bay during the LGM and are indicative of relatively fast ice flow (cf.
668 King et al., 2009). The absence of recessional moraines, which are commonly
669 observed in glaciated areas (e.g. Landvik, 1994; Ottesen et al., 2005; Ottesen &
670 Dowdeswell, 2009; Dowdeswell et al., 2010; Hogan et al., 2010), suggests that
671 retreat was so rapid that there was insufficient time for recessional moraines to
672 form, or that retreat occurred as a floating ice shelf. An interpretation of rapid
673 retreat is supported by the reconstructed high retreat rates from the continental
674 shelf and from Disko Bay (discussed in section 5.3 below).

675 In terms of the Holocene sedimentary environments and the associated depo-
676 sitional processes in front of Jakobshavn Isbræ and adjacent GIS outlet glaciers,
677 we identify four main processes: (1) suspension settling of glacimarine muds
678 from meltwater and the water column, (2) meltout of coarser grains from ice-
679 bergs and sea ice (3) sediment gravity flows, reworking both fine- and coarse-
680 grained deposits down the slopes of submarine basins, and (4) ploughing of
681 sediments by grounded iceberg keels. Meltwater-derived sedimentation is the
682 dominant process, as indicated by the exceptionally well-sorted muds (usually
683 >95% of the muds have a grain size $<63\ \mu$) and their overall strong terrestrial
684 signal.

685 5.2. Timing of ice retreat and deglacial ice sheet dynamics

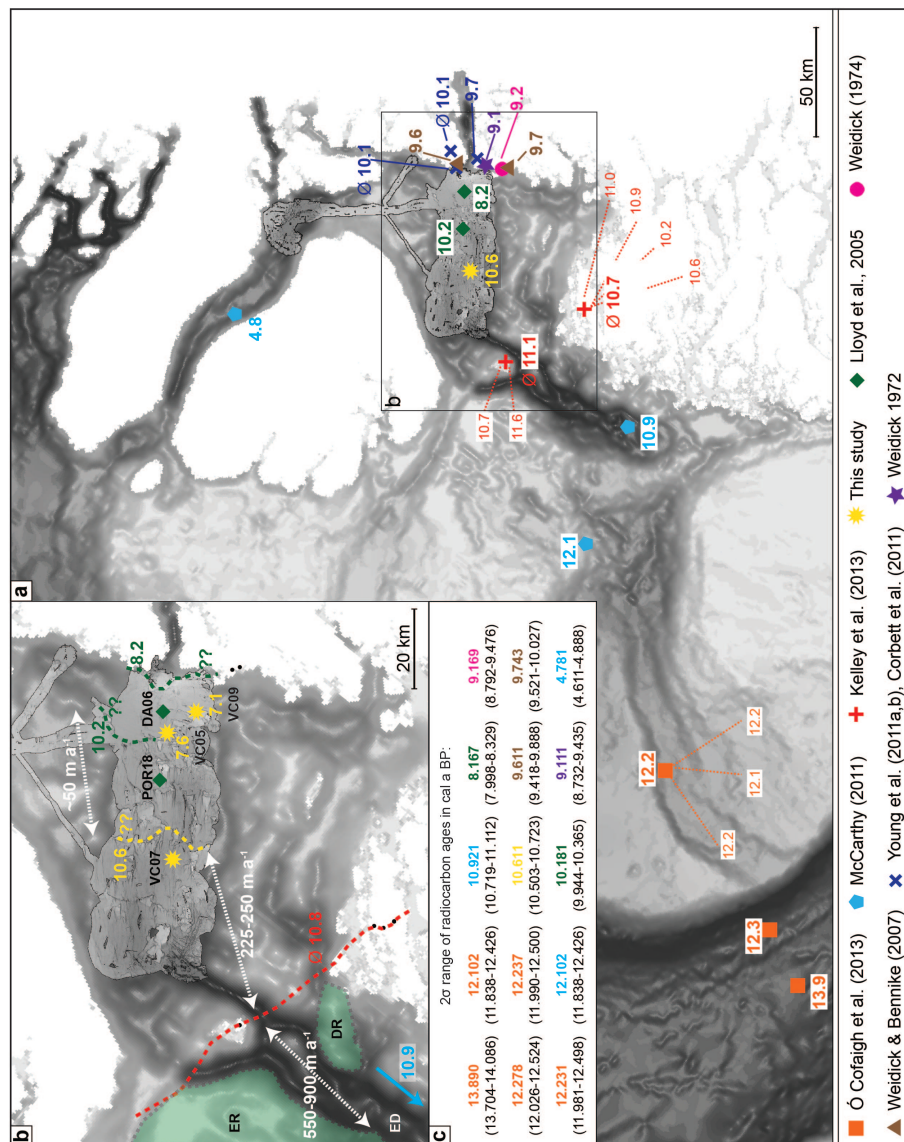
686 In order to reconstruct a retreat chronology for Jakobshavn Isbræ, we integrate
687 the radiocarbon dates presented in this study with previously published data
688 from both marine and terrestrial settings in West Greenland (Fig. 10). An
689 extended Jakobshavn Isbræ and its adjacent GIS outlets flowed through Disko
690 Bay and through the Vaigat Strait onto the continental shelf during the Last
691 Glacial Maximum (cf. Ó Cofaigh et al., 2013; Dowdeswell et al., 2014). The
692 ancestral Jakobshavn Isbræ commenced retreat from the outer continental shelf
693 around 13.8 ka BP or before, and underwent a short-lived re-advance during
694 the Younger Dryas on the shelf west of Disko Bay, at ~ 12.3 – 12.0 ka BP (re-
695 calibrated from Ó Cofaigh et al., 2013). Deglaciation of the inner continental
696 shelf and the western parts of Disko Bay was underway by 10.9 and 10.8 ka
697 BP, respectively (Fig. 10; McCarthy, 2011; Hogan et al., 2012; Kelley et al.,
698 2013). Our date of 10.6 ka BP from relatively proximal glacimarine sediments
699 in central Disko Bay serves as a minimum age for deglaciation of the central
700 bay and demonstrates that the ice margin had retreated to a position east of
701 VC07 by this time. The 2.8 m-thick sequence of proximal sediments at the base
702 of VC07 must have been deposited before this date and could indicate that the
703 ice margin retreated very slowly (see also section 5.3 below) or paused close
704 to the core site of VC07. A radiocarbon date published by Lloyd et al. (2005)
705 from core POR18 showed that the ice margin was located at or somewhere
706 within 6 km west of Isfjeldsbanken at 10.2 ka BP (re-calibrated; Fig. 10), and
707 is in good agreement with findings from Young et al. (2011a), who place the
708 ice margin at or close to Isfjeldsbanken at 10.2 ka BP. Sub-bottom profiler
709 data presented by Hogan et al. (2011) suggest a prolonged still-stand of the ice
710 margin at Isfjeldsbanken, which led to the accumulation of thick sedimentary
711 basin infills immediately west of the sill. The radiocarbon dates from thick
712 ice-proximal sedimentary sequences from inner Disko Bay (VC09 and VC05,
713 this study; DA00-06, Lloyd et al., 2005; Hogan et al., 2011) further support
714 this. Two re-advances of Jakobshavn Isbræ occurred in response to climatic

cooling events around 9.3 and 8.2 ka BP (e.g. Young et al., 2013), the latter of which may be reflected in the transition from more distal (lithofacies LD2a) to more proximal glacimarine muds (LD2b) in VC09. The change from relatively proximal glacimarine (LD2b) to predominantly hemipelagic sediments (LD2c) in VC05 and VC09 implies that Jakobshavn Isbræ had retreated into Isfjorden sometime around 7.6–7.1 ka BP. This is consistent with work from Lloyd et al. (2005) and Hogan et al. (2011), who concluded that the ice stream retreated from Isfjeldsbanken into Isfjorden at c. 7.9–7.8 ka BP. Although Corbett et al. (2011) and Young et al. (2011b,a, 2013) suggested slightly earlier deglaciation of the coastal areas in eastern Disko Bay, it is possible that Jakobshavn Isbræ remained grounded at Isfjeldsbanken longer than the surrounding ice masses. During subsequent retreat the ice margin withdrew to a position behind that of the present terminus, where it remained throughout the mid-Holocene (Young et al., 2011a, 2013; Kelley et al., 2013), before it re-advanced westwards to its present position after 2.2 ka BP (cf. Weidick & Bennike, 2007).

In the Vaigat Strait, the chronology of deglaciation is less clear, reflecting a lack of data from this region but also the very thick post-glacial sediment cover, which makes it difficult to obtain radiocarbon ages relating to ice retreat through the strait. In fact, to our knowledge, the oldest date of 4.8 cal ka BP (re-calibrated) was obtained at a depth of 435 cm from a core from the central Vaigat Strait (Fig. 10; McCarthy, 2011), and is regarded as a minimum age for ice retreat.

5.3. Retreat rates

From the above constraints on the timing of ice retreat, minimum rates of retreat can be estimated for the ice sheet outlet glaciers, which retreated relatively quickly across the continental shelf, accelerated through Egedesminde Dyb, slowed slightly through western Disko Bay, and significantly slowed through the eastern bay. Note that the rates presented here are average rates, based on the assumption of linear ice retreat. The geomorphological and lithological



744 evidence presented in this study suggests that the margin of Jakobshavn Isbræ
 745 was grounded during retreat. This is based on the implication of meltwater-
 746 dominated sedimentation with moderate input from icebergs and sea ice in the
 747 cores from Disko Bay and the Vaigat Strait, as sediment facies associated with
 748 ice shelves tend to be diamictic and coarse-grained close to the grounding line,
 749 and, in the case of the more distal sub-ice shelf sediments, tend to lack IRD (cf.
 750 e.g. Anderson et al., 1991; Powell et al., 1996; Domack et al., 1999; Kilfeather
 751 et al., 2011). Furthermore, the presence of a large submarine channel (C1, see
 752 section 4.1 and Figs. 3, 4) in the central bay is thought to support a grounded
 753 ice margin, because its depth, shape, and the close association with a trans-
 754 verse bedrock ridge seem consistent with subglacial meltwater excavation at the
 755 grounding line of a stagnating ice margin over an extended period of time. Nev-
 756 ertheless, the absence of recessional features on the seafloor in Disko Bay and
 757 the periodic appearance of coarse-grained diamicts (LD1 in VC03, VC04, and
 758 VC08) may reflect transient decoupling of the ice stream from its underlying
 759 bed. Indeed, a lightly grounded and relatively thin ice margin with predom-
 760 inantly hydrostatic support was already proposed by Hofmann et al. (2016),
 761 who further suggested that the ice stream may have grounded intermittently
 762 at bedrock highs. It is therefore possible, that, while our data mainly imply
 763 grounded ice, the ice stream experienced occasional periods of ungrounding, at

Figure 10 (*preceding page*): a) Summary of radiocarbon (median) and ^{10}Be dates available from marine organisms on the continental shelf and Disko Bay, and from bedrock from adjacent land areas in (cal) ka BP. Radiocarbon dates from previously published studies were re-calibrated using a ΔR of 140 ± 25 years (Lloyd et al., 2005). b) Zoom-in on the study area according to the rectangle indicated in a). Stippled red, yellow and green lines show an estimation of where the ice front position could have been based on the dates from boulders/bulk sediment (red; Kelley et al., 2013) and sediment cores (yellow for this study and green for Lloyd et al., 2005). White arrows and numbers imply possible retreat rates. Green shaded areas show Egedesminde Ridge (ER). ED = Egedesminde Dyb, DR = Disko Gneiss Ridge (cf. Hofmann et al., 2016). c) Summary of the 2σ ranges obtained from re-calibration of the radiocarbon ages.

764 least locally.

765 The ancestral Jakobshavn Isbræ retreated at rates between 22–275 m a⁻¹
766 across the continental shelf after its Younger Dryas re-advance around 12.3–
767 12.0 ka BP (Ó Cofaigh et al., 2013). A retreat of up to 90 km between c. 10.9
768 ka BP (re-calibrated radiocarbon date from McCarthy, 2011) and 10.8 ka BP
769 (¹⁰Be date from Kelley et al., 2013, Fig. 10) implies accelerated retreat through
770 Egedesminde Dyb at rates between 550 and 900 m a⁻¹. Although these rates
771 are much higher than those for the continental shelf, more extensive calving and
772 thus faster ice retreat has often been linked to bathymetric overdeepening (e.g.
773 Meier & Post, 1987; Seramur et al., 1997; Oerlemans & Nick, 2006; Benn et al.,
774 2007; Ó Cofaigh, 1998; Kehrl et al., 2011). Furthermore, Egedesminde Ridge
775 west of the trough (Fig. 10), could have served as a pinning point during ice
776 retreat (Hofmann et al., 2016). Once the ice stream detached from this ridge,
777 fast retreat would have occurred due to increased glacier bottom melting caused
778 by the inflow of warm Atlantic water into the trough (cf. Andersen, 1981; Lloyd
779 et al., 2005; Lloyd, 2006; Holland et al., 2008). The subsequent retreat across
780 outer Disko Bay occurred over a minimum distance of 45 km between ~10.8
781 and 10.6 ka BP to a position east of VC07, suggesting minimum rates between
782 225 and 250 m a⁻¹ from the western to the central bay (Fig. 10). This shows
783 that retreat either slowed, or that Jakobshavn Isbræ temporarily paused upon
784 entering Disko Bay, which is likely, given the sudden shoaling from >1100 to 400
785 m water depth at the eastern end of Egedesminde Dyb and a sudden widening
786 of the retreat basin (e.g. Oerlemans & Nick, 2006; Benn et al., 2007). Retreat
787 was even slower through the eastern parts of the bay, as shown by dates of
788 10.6 ka BP from proximal glacimarine sediments in VC07 and 10.2 ka BP from
789 similar sediments in POR18, which indicate a retreat of approximately 20 km
790 at a rate of ~50 m a⁻¹ (Fig. 10; Lloyd et al., 2005, this study). Similar rates
791 were obtained from the cosmogenic dates in the area, thus supporting slow or
792 intermittent retreat through eastern Disko Bay (Kelley et al., 2013).

793 5.4. Comparison between West Greenland and other glacima- 794 rine environments

795 Suspension settling, ice rafting, sediment gravity flows, and iceberg ploughing
796 were identified as the key sedimentary processes during deglaciation of Disko
797 Bay and the Vaigat Strait. Although these four processes reflect those com-
798 monly observed in high-Arctic fjord environments (e.g. Elverhøi et al., 1983;
799 Powell & Molnia, 1989; Andrews et al., 1994; Syvitski et al., 1996; Dowdeswell
800 et al., 1998; Ó Cofaigh & Dowdeswell, 2001; Forwick et al., 2010), the variable
801 magnitude of each of these has important implications for our understanding
802 of glacial marine sedimentation. Thus far, depositional environments in front of
803 tidewater glaciers have been categorised according to climatic and glaciological
804 regime (Dowdeswell et al., 1998). Southeast Alaska forms the warmer end of the
805 spectrum, with predominantly fine-grained mud deposited from glacial meltwa-
806 ter (Powell & Molnia, 1989; Cowan & Powell, 1991). Antarctica forms the other
807 extreme, defined as a polar and climatically severe setting, where sedimentation
808 occurs mainly at the grounding line (e.g. Domack et al., 1999; Powell et al.,
809 1996; Ashley & Smith, 2000). Fjords around Svalbard and Baffin Island are in
810 between these two end members (Dowdeswell et al., 1998) with high amounts
811 of meltwater-derived muds close to the glacier fronts (e.g. Elverhøi et al., 1983;
812 Gilbert, 1983; Gilbert et al., 1990; Forwick et al., 2010; Streuff et al., 2015), but
813 increasing amounts of ice-rafted material towards ice-distal areas (e.g. Elverhøi
814 et al., 1983; Forwick & Vorren, 2009; Kempf et al., 2013). East Greenland
815 was initially defined as an environment with low meltwater availability, where
816 sedimentation is dominated by iceberg-rafting and meltout from sea ice (e.g.
817 Marienfeld, 1991; Syvitski et al., 1996; Dowdeswell et al., 1993, 1994, 1998).
818 However, subsequent work by Smith & Andrews (2000) and Ó Cofaigh et al.
819 (2001) showed that large amounts of fine-grained stratified sediments in prox-
820 imal areas of East Greenland fjords record sedimentation predominantly from
821 meltwater, and that deposition of IRD only becomes important in more ice-distal
822 environments. Large amounts of silt and clay derived from meltwater were also

823 observed in other East Greenland fjords (Andrews et al., 1994). Accordingly,
824 Ó Cofaigh et al. (2001) proposed that glacimarine sedimentary processes can be
825 very similar despite different climatic, glaciological and oceanographic settings,
826 and that their variability may rather be a consequence, at least in part, of local
827 controls, such as distance to the ice margin.

828 There has been limited research investigating glacimarine sedimentary pro-
829 cesses in West Greenland and it has not been considered in the spectrum out-
830 lined above, perhaps due to the only recently emerging data (e.g. McCarthy,
831 2011; Jennings et al., 2013; Ó Cofaigh et al., 2013; Dowdeswell et al., 2014;
832 Hogan et al., 2016; Sheldon et al., 2016). The abundance of meltwater-derived
833 sediments in the cores from Disko Bay and the Vaigat Strait emphasise the im-
834 portance of meltwater sedimentation in proximal areas of GIS outlets here and
835 suggest that the ice-proximal sedimentary processes in West Greenland are com-
836 parable with those from warmer settings like Svalbard and Alaska (e.g. Powell
837 & Molnia, 1989; Cowan & Powell, 1991; Cai et al., 1997; Forwick & Vorren,
838 2009; Forwick et al., 2010; Streuff et al., 2015). Considering the nearly identical
839 mean annual air temperatures and annual precipitation between Svalbard and
840 West Greenland and that both are influenced by relatively warm and saline At-
841 lantic water, similar depositional processes may not be surprising. Similarity in
842 sedimentary processes also suggests that in terms of depositional environment
843 Disko Bay acts more like a fjord than a marine embayment on the continental
844 shelf. However, the increasingly hemipelagic and diamictic sediments and the
845 associated reduction in meltwater flux in the distal areas of Disko Bay (VC08
846 and VC07) are different from Svalbard and Alaska, where sedimentation from
847 meltwater remains the dominant process throughout the entire glacimarine set-
848 ting (Görlich et al., 1987; Boulton, 1990; Streuff et al., 2015). This strongly
849 implies that glacimarine processes and their associated facies are not simply
850 a function of climate. In fact, Disko Bay appears to be more similar to the
851 glacimarine depositional environments of East Greenland fjords, which is no-
852 table given the classification of East Greenland as a polar, meltwater-restricted
853 glacimarine environment, and the extensive sea ice in most of its fjords. The

comparatively low SARs in Disko Bay with respect to those in East Greenland fjords may be related to differences in the availability of meltwater or glacial debris, or to the different fjord morphology compared to Spitsbergen and East Greenland fjords (wide open bay vs. narrow constricted fjords). It follows that even within geographically constrained areas glacimarine sedimentary processes and their magnitude can vary significantly over distance and time. We conclude that variability between meltwater-dominated and iceberg-dominated glacimarine sedimentation is not necessarily related only to climate and glaciology but is also dependent on local factors including distance to the ice margin, seafloor topography and glacier size (cf. Ó Cofaigh et al., 2001).

6. Conclusions

Lithological data integrated with swath bathymetry and TOPAS sub-bottom profiler data provide new insights into the Holocene glacimarine sedimentary processes in Disko Bay and the Vaigat Strait in West Greenland. Vibrocores comprise diamict, (diffusely) stratified mud, massive mud with sharp-based sand layers, IRD-rich massive mud, and massive bioturbated muds. These facies show that suspension settling of fine-grained sediment from turbid meltwater plumes and the water column, sediment gravity flows, and iceberg rafting and ploughing were the dominant sedimentary processes during and following ice retreat, with meltwater sedimentation dominant in ice-proximal areas, and hemipelagic suspension settling and IRD-rainout from icebergs dominant in distal areas.

Our findings show that despite similar climate and oceanography glacimarine sedimentary processes differ between Svalbard and West Greenland, but are similar between East and West Greenland in spite of different oceanographic conditions. This confirms that such processes vary more as a function of local controls such as distance from the ice margin and geomorphological setting rather than climate and geographic location. Radiocarbon dates provide the basis for estimated SARs between 0.1 and 1.7 cm a⁻¹ in proximal areas, and ~0.007–0.05 cm a⁻¹ in distal areas, which are lower than SARs documented for

883 East Greenland. The radiocarbon dates further constrain the retreat dynamics
884 of Jakobshavn Isbræ during deglaciation. Streamlined glacial landforms, includ-
885 ing crag-and-tails and glacial lineations, record the former flow of an expanded
886 Jakobshavn Isbræ and adjacent GIS outlets through Disko Bay and the Vaigat
887 Strait towards the adjoining continental shelf. During deglaciation, retreat was
888 relatively fast across the continental shelf ($22\text{--}250\text{ m a}^{-1}$), through Egedesminde
889 Dyb ($\sim 550\text{--}900\text{ m a}^{-1}$), and the western parts of Disko Bay ($\sim 225\text{--}250\text{ m a}^{-1}$),
890 all of which were deglaciated before 10.6 ka BP. Subsequent retreat through
891 eastern Disko Bay was much slower ($\sim 50\text{ m a}^{-1}$), and likely interrupted by at
892 least one still-stand due to pinning of the grounded glacier margin on submarine
893 bedrock ridges. The ice margin paused again at Isfjeldsbanken before retreating
894 into Isfjorden. Around 7.6–7.1 ka BP the ice margin had probably retreated
895 far back into Isfjorden, as at this point sediment delivery to the core sites from
896 meltwater plumes became significantly reduced. The variable retreat rates and
897 sedimentary facies we document here underscore the importance of local mor-
898 phology and glacier proximity for the palaeo-retreat dynamics and associated
899 glacialmarine sedimentary processes of marine-terminating Greenland Ice Sheet
900 outlet glaciers.

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